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THOROLF VOGT

Noen minneord.

Med Thorolf Vogt er en av Norges ledende geologer, og en av de mest allsidige, gått bort.

Han var født i Vang i Hedemark 7. juni 1888, sønn av Johan Herman Lie Vogt og Martha Johanne Abigael Kinck. Etter artium i 1906 begynte hans vitenskapelige løpebane nesten med det samme: i Norsk geologisk tidsskrift publiserte han et arbeide allerede i 1908. Det fulgte en lang rekke andre arbeider, og i 1928 tok han doktorgraden ved Universitetet i Oslo på sin store avhandling om Sulitjelmafeltets geologi og petrografi. Han gjorde studiereiser til Wien 1910—11 og til Göttingen 1913. Videre reiste han flere ganger i Sverige, Danmark, Storbritannia, U.S.A. og andre land. I 1925 og 1928 ledet han ekspedisjoner til Spitsbergen, og i 1931 en ekspedisjon til det sydøstlige Grønnland. Ved Norges geologiske undersøkelse var han assistent 1909—14, og statsgeolog 1914—29. Ved Universitetet i Oslo var han universitetsstipendiat 1915—23. I 1929 ble han utnevnt til professor i mineralogi og geologi ved Norges tekniske høgskole. I dette embete ble han til han plutselig døde den 8. desember 1958. Han hadde da nådd aldersgrensen, men var på oppfordring blitt stående i stillingen.

Vogt publiserte over 100 vitenskapelige arbeider, deriblant en rekke meget betydningsfulle. Det er ikke meningen å gå i detaljer her,— det henvises til den etterfølgende fullstendige bibliografi. Her skal det bare minnes om noen hovedlinjer i Vogts vitenskapelige virksomhet.

Til å begynne med arbeidet han vesentlig innen mineralogien, og han publiserte en rekke utmerkede avhandlinger på dette felt. Til tross for at han senere konsentrerte seg om geologiske og petrografiske undersøkelser, bevarte han alltid en levende interesse for mineralogien; hans aller siste, ennå ikke trykte, publikasjon — et hornblende-arbeide — er et av bevisene for dette. Som statsgeolog arbeidet han i forskjellige deler av Norge, og hans bidrag til Norges regionale geologi er meget viktige. Her må nevnes hans betydningsfulle studier over stratigrafien i sparagmitområdet ved Mjøsa. Men hans viktigste arbeidsfelt gjennom mange år var den kaledonske fjellkjede, dens tektonikk, bergarter og malmer, først og fremst i Nord-Norge, men også i Trondheimsfeltet. Hans betydeligste publikasjon på dette område er den store avhandlingen om Sulitjelmafeltet, en bok på 560 sider med mange plansjer, geologiske kart og profiler. Behandlingen av stratigrafi, tektonikk og petrografi, som selvsagt er meget inngående, inneholder også teoretiske avsnitt av generell interesse. Særlig kjent, og ofte sitert av petrografer, er avsnittet om anvendelsen av mineralfasieslæren på feltets metamorfe bergarter, et petrologisk viktig bidrag. Denne boken er egentlig bare første del av en avhandling som også skulle omfatte Sulitjelmafeltets malmer. Den andre delen ble aldri publisert, og det foreligger bare mindre arbeider om kaledonske malmforekomster fra Vogts hånd, således er innlegg i diskusjonen om disse malmenes genesis. En rekke små avhandlinger handler om geokjemisk og geobotanisk malmleting. Et overveiende stratigrafisk arbeide fra et kaledonsk område er den viktige avhandling om Hølonda-Horg-distriktet i Trondheimsfeltet. Kaledonidenes store linjer og utviklingshistorie interesserte Vogt gjennom lang tid, og han diskuterte disse problemene på grunnlag av sine rike erfaringer i flere publikasjoner. Et par av hans siste arbeider behandler disse ting, og her er hans betraktninger ikke begrenset til norsk territorium, men tar også med den skotske del av kaledonidene. — Leilighetsvis har Vogt behandlet en rekke andre emner, som isostasi, glasialgeologi og meteoritter.

Vogt var medlem av Det norske videnskapsakademi i Oslo, og av Det Kgl. Norske Videnskabers Selskab i Trondheim, hvor han i senere år var visepreses og siden preses. I Norsk geologisk forening var han formann i 1919 og 1928.

Vogt var kommandør av St. Olavs orden.

Ivar Oftedal.

Kronologisk fortegnelse over professor dr. philos Thorolf Vogt's trykte vitenskapelige arbeider.

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Norsk Geologisk Tidsskrifts forkortelser:

Abbreviations:

G.F.F.: Geologiska Föreningens i Stockholm Förhandlingar.
N.G.T.: Norsk Geologisk Tidsskrift.
N.G.U.: Norges geologiske undersøkelse.
V.A.O.: Det Norske Vitenskapsakademi i Oslo.
V.S.S.: Det Kgl. Norske Videnskabers Selskab.

De øvrige forkortelser er tatt fra:

Other abbreviations according to:

«World list of scientific periodicals published in the years 1900—1950. London 1952.»

THE DOUBLE CARDINAL PROCESS IN THE OLDEST STROPHOMENIDS

By

NILS SPJELDNÆS

Abstract: Some of the oldest Strophomenidae with bifid cardinal processes are described. It is suggested that at least in some of them, the bifid cardinal processes were formed by partial neoteny. The "chilidal plates" of the Sowerbyellinae are regarded as analogous with the socket plates of the Strophomenidae. Further ontogenetical studies are necessary to tell if the family Strophomenidae is monophyletic.

Introduction

The classification of the strophomenid brachiopods is based mainly on the development of the cardinal processes (cf. WILLIAMS 1953a). The Plectambonitidae have simple, orthoid, cardinal processes, the Leptestiidae have complex ones, and the Strophomenacea have bifid cardinal processes.

The author's studies of Norwegian Middle Ordovician Strophomenids seemed to indicate the possibility that the development of the bifid cardinal processes in the Strophomenacea was not monophyletic. (cf. SPJELDNÆS 1957, pp. 194—196). It seemed therefore desirable to study the development of the cardinal processes in the oldest members of the Strophomenacea, in order to see if these species might throw some light on this important problem. A review of the litterature showed that relatively few Lower Ordovician Strophomenidae are known (the Christianiidae were described by SPJELDNÆS 1957). The genus *Kirkina* SALMON 1942, with the only species *K. milliardensis*

SALMON is among the oldest known, but it is not well known, and no material of this species was at the author's disposal. *Rafinesquina*(?) *macallumi* REED 1917 is present in the author's collections, but the age of this species is not definitely known. The present note is therefore mainly based on some specimens of *Rafinesquina*(?) *llandeiloensis* (DAVIDSON), from Llan Mill, near Narberth. All the specimens are found in one slab, which was borrowed from the Geological Survey and Museum, London.

This note is an appendix to the author's studies on the Middle Ordovician Strophomenids of the Oslo Region (SPJELDNÆS, 1957). He is deeply indebted to prof. L. Størmer and prof. Alwyn Williams, Belfast, for inspiring discussions in connection with the problems dealt with in this paper. Furthermore, the author wishes to express his thanks to the Faculty Photographer, Miss B. Mauritz, who has taken the photographs.

The oldest Strophomenacea

Genus *Kirkina* SALMON 1942.

Type species: K. milliardensis SALMON 1942 (pp. 599—600, text-fig. 7, pl. 87, figs. 34—35. Cf. also COOPER 1956, pp. 866, 160.)

Remarks: This genus is not well known. The only species is reported to have undifferentiated sculpture and very small pseudopunctae, which was supposed to separate it from *Rafinesquina*. Only the dorsal interior is known from the type species, and it does not differ from that of *Rafinesquina* in any important features. Until more material has been studied, it is difficult to estimate the validity of this genus, it might be only a subgenus of *Rafinesquina*, such as *Kjaerina* and *Hedstroemina*.

The stratigraphical range of the type species was unknown to SALMON, but according to HINTZE (1952, pp. 22—23) it is found in several localities in zone "N" in Utah and Nevada. ROSS (1951, p. 33) records *Didymograptus bifidus* from zone "M" in the same area. COOPER (1956, pp. 160, 866) includes these beds in the Lehman formation of the Whiterock Stage. The upper part of this stage probably corresponds to the upper part of the Llandvirn, or the lowest Llandeilo of Europe.

Rafinesquina(?) *macallumi* Reed 1917.

Pl. 1, figs. 8—9.

1917 *Rafinesquina expansa* (Sow.) var. nov. *macallumi* — Reed, pp. 865—866, pl. 11, figs. 17—20.

Age and distribution: Reed recorded this species from a number of localities in the Girvan District, all from the Stinchar Limestone (s. l.). Only the exteriors were known to him. The present material comes from Minuntion, in the *confinis*-grits, and probably at the same horizon at Bougang. The age of the beds are not known exactly, but they are below the Stinchar Limestone proper, which is succeded by the Didymograptus Shale, with graptolites possibly belonging to the *Nemagraptus gracilis* zone. (According to Pringle 1948, pp. 18—20, this graptolite fauna perhaps has a high position in the Glenkiln Shales. The presence of *Didymograptus superstes* and *Glossograptus hincksi* indicate, however, that the fauna is intermediate between the *gracilis* and the *teretiusculus* zones.).

The trilobites found by the author together with *R.*(?) *macallumi* give no direct information on the stratigraphy. Among the brachiopods, the author found *Sowerbyella* cf. *llandeiloensis*. This species seems to indicate an Upper Llandvirn or Lower Llandeilo age of the beds, but it might easily be confused with young specimens of other species. The age of the *confinis*-grits are therefore not definitively determined, but it is probably below the Caradoc.

Description: The size, outline and sculpture of the present specimens are identical with those figured by Reed. The sculpture is differentiated, with 3—7 striae between each rib. The new ribs are all formed along the same concentric line, such as in *Oslomena* (cf. Spjeldnæs 1957, p. 164). The valves are moderately convex, and without geniculation. In some large specimens, the margin is abruptly deflected.

The dorsal interior is similar to that of *Rafinesquina*(?) *llandeiloensis*, except that the socket plates are less defined, and that the cardinal processes are longer and thinner. These features might be due to the different size of the specimens.

A young dorsal interior from Bougang showed some other features, having well defined socket plates, no anchor-shaped platform, and semiparallel cardinal processes. Unfortunately the specimen is slightly distorted, so that it is unsymmetrical. It seems as if the cardinal processes are placed on both sides of a short, deep cavity. This specimen resembles the figured specimens (pl. 1, figs. 2, 3, 5) of the young valves of *R.*(?) *llandeiloensis*.

The ventral interior is also like that of *R.*(?) *llandeiloensis*, with bundles of fine septa instead of distinct muscle impressions. The teeth are larger in the present species, possibly because the specimens are larger, and have no marginal thickening.

Remarks: As mentioned above, the difference between this species and *R.*(?) *llandeiloensis* is rather small, only the sculpture is different. It is therefore possible that further studies on more material will show that *macallumi* is only a subspecies of *llandeiloensis*.

Rafinesquina(?) *llandeiloensis* (DAVIDSON 1866—71).

Pl. 1, figs. 1—7.

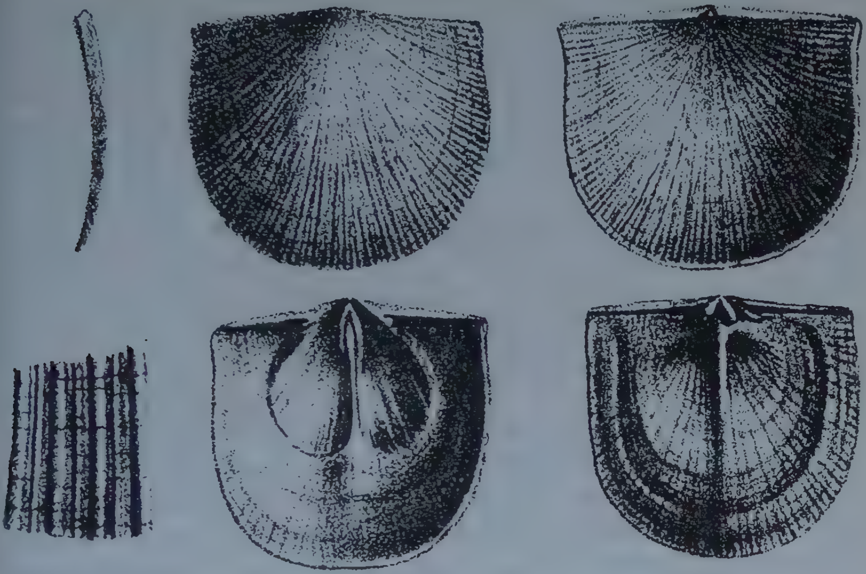
1866—71 *Strophomena compressa* (SOW.) var. *llandeiloensis* — DAVIDSON, pp. 316—317, pl. 46, figs. 11—14.

1953 *Rafinesquina llandeiloensis* (DAV.) — WILLIAMS, pp. 184—193 (not figured).

Age and distribution: This species is recorded from several horizons and localities in the Llandeilo District by WILLIAMS (1953, pp. 184—193), ranging from the Ffairfach Group (Upper Llandvirn) to the Lower Llandeilo. Judging from the associated fauna (*Marrolithus inflatus incipiens* WILLIAMS, *Flexicalymene cambrensis*, *Horderleyella* sp., *Dalmanella* cf. *parva*, *Tallinnella complicata* and other ostracods, *Sowerbyella antiquata llandeiloensis* and bryozoans) the present material comes from the zone of *Marrolithus inflatus incipiens*, the uppermost part of the Llandvirn. The material comes from Llan Mill, near Narberth, Carmarthenshire, which is the type locality for *R.*(?) *llandeiloensis*. The specimens are all found on one slab, which belongs to the Geological Survey and Museum, London. About 20 interiors are found on the slab, more or less fragmentary.

Description: The sculpture is variably differentiated. In some specimens only every second or third rib is slightly stronger than the others, and in others there are from two to eight striae between each rib.

The valves are moderately convex, the dorsal one geniculating



Textfig. 1. *Rafinesquina* (?) *llandeiloensis* (Davidson). The original figures from Davidson (1866—71) pl. 46, figs. 11, 13, 14. The specimens are from Llan Mill, Carmarthenshire. About 1 ½ times the original size.

dorsally in large specimens. The ventral valves are generally somewhat more convex than the dorsal ones, and have a marginal thickening. The areas are narrow, the delthyrium and notothyrium wide. A small, strongly curved, thin pseudodeltidium is present in large specimens, and a low chilidium is always present even in the larval valves.

The ventral interior shows slightly developed muscle impressions which seem to consist of bundles of fine septa, resembling those of *Grorudia* and *Paleostrophomena*. Only in one of the specimens present are the muscle impressions complete, such as in that figured by Davidson (textfig. 1.). The teeth plates are strong, and continuous with the elevated edges of the delthyrium (*e*, pl. 1, fig. 1). Even small valves are generally very thick, and the marginal deposits of calcareous tissue are conspicuous (*mt.*, pl. 1, fig. 1).

The dorsal interior of adult specimens shows two long, diverging cardinal processes resting on a high, anchor-shaped platform. The muscle attachment surfaces are inclined steeply towards the notothyrium. The socket plates, forming the lateral branches of the "anchor", are strongly developed. The sockets are very shallow, scarcely discernable except in young specimens (pl. 1, fig. 2, *sc*).

On the anterior side of the socket plates, there are in some cases a number (3—4) of crenulations, resembling those found in the earliest Stropheodontids. They are, however, not found in all specimens, and might in some cases be casts of the coarse-grained matrix. In the figured specimen (pl. 1, fig. 4, *cr*) they are very strong. Such "pseudocrenulations" of the teeth and socket plates are known from a number of Strophomenid species (SPJELDNÆS 1957, pp. 20—32, 190—191), and do not necessarily mean that this species belongs to the Stropheodontidae, even if the resemblance with *Stropheodonta* (*Eostropheodonta*) *williamsi* SPJELDNÆS (1957, pl. 12, figs. 3—5), the oldest real Stropheodontid known at present, is striking.

The chilidium is low, but the chilidal plates (pl. 1, fig. 2 *ch*) are high, thin and sharply delimited from the socket plates. In younger Strophomenid species, it is generally ancylosed to the socket plates and can only be distinguished by its histological structure. (cf. SPJELDNÆS 1957, textfig. 40—41).

The young dorsal interiors show additional features (pl. 1, figs. 2, 3, 5). The notothyrial platform is absent, the cardinal processes are resting directly on the notothyrial floor, and the muscle attachment surfaces are parallel with the interior surface of the valves. A distinct, horizontal slit is found between the cardinal processes (*sl*, pl. 1, fig. 2). The socket plates are distinctly developed, and the cardinal processes are fused to their proximal ends. In the smallest specimens, (pl. 1, fig. 5) they can scarcely be discriminated from them. The chilidium is comparatively better developed in the small specimens.

The trend outlined in the larval development of the valves of *R.* (?) *llandeiloensis* seems to be the same as that found in *Christiania holtedahli* and *Kjerulfina foliovalve* described from the Middle Ordovician of the Oslo Region (SPJELDNÆS 1957, pp. 51—52, 117—120, 155—156).

The sockets and the area just outside them are covered with tissue with very large pseudopuncta (in the adult specimens). In old specimens this tissue is concentrated at the socket plates and perhaps the "pseudocrenulations" originated from these coarse pseudopuncta. This concentration of large pseudopuncta is found also in the

Rafinesquina sardesoni-group of SALMON (1942, p. 572, textfig. 5 b). In *Oepikina*, the pseudopuncta are considerably larger in the central part of the valve, near the hinge line, than in the marginal parts. In this case, however, the large pseudopuncta are not restricted to the sockets and the area just outside them, such as in the *Rafinesquina*-species in question.

Neither in the adults nor in the larval valves are there any traces of a medial cardinal process.

Oslomena (?) sp.

Pl. 1, fig. 10.

1917 *Rafinesquina* (*Playfairia*) *semiglobosina* (DAVIDSON) parte. — REED, pp. 869—870, non pl. 12, figs. 13—20.

In the material from Minuntion, Girvan District, Scotland, there is a ventral interior belonging to this species. The valve is strongly convex, with a very slight geniculation, and with flat cardinal angles.

The interior has small, but well defined muscle impressions, which resemble those found in the genus *Oslomena* SPJELDNÆS (1957, pp. 161, 164, textfig. 34, p. 141.), and is different from those of *Rafinesquina* both in size and shape.

This species is mentioned here because it shows that the *Oslomena*-type of Strophomenids occurs together with the *Rafinesquina*-type in beds of this age.

TAXONOMY

It is outside the scope of this paper to discuss the taxonomy of the earliest Strophomenidae, and therefore it is only pointed out that the earliest species (*llandeiloensis* and *macallumi*) probably do not belong to *Rafinesquina* as it is defined now (SALMON 1942, COOPER 1956). It is restricted to the Upper Ordovician and the uppermost part of the Middle Ordovician, and the species described here are from the lowest part of the Middle Ordovician. According to BANCROFT (1929, p. 41—42) a similar species from the Hoar Edge Grit in Shropshire belongs to *Rafinesquina*, and not to *Kjaerina* or *Hedstroemina*. The genus *Kirkina* SALMON is unfortunately not well known, and further studies might show that the present species can be referred to it.

Macrocoelia plebeia COOPER (1956, p. 899, pl. 121G, 233B) is also similar in some respects, but both *plebeia* and the species described here differ considerably from the rest of the species in that genus. At present it seems convenient to refer all these species to *Rafinesquina* (?).

The cardinal processes

From what is described above, and from the results obtained by SPJELDNÆS (1957, pp. 23—26, 50—52, 194—196) the following conclusions might be summarized:

1) No signs of a trifid, Sowerbyellin cardinal process is observed in the ontogeny of the typical Strophomenidae.

2) In *Oepikina* and some species of *Christiania*, a modification of the Sowerbyellin cardinal process is found in the adults.

3) In all larval specimens belonging to the Strophomenidae there is a distinct groove beween the cardinal processes. This groove is supposed to be the pedicle groove.

4) A striking difference is found between the ontogenetic and the phylogenetic development of the cardinal processes in *Christiania*.

The ontogenetic development of all known protrematous brachiopods seems to indicate that the pedicle was impressed in both valves in the youngest larval stages, and was later transferred to the ventral valve only. In most brachiopod groups (Orthids, Clitambonitids, Plectambonitids) the cardinal processes developed after the pedicle was transferred to the ventral valve, or after it was reduced or lost. In the Strophomenidae, the bifid cardinal processes were probably formed when the pedicle was still impressed in the dorsal valve also, and a median cardinal process could not be formed.

This type of development seems to be an example of partial neoteny (merostasis), since one larval feature (the pedicle impressed in the dorsal valve) is retained until a comparatively late stage in ontogeny. STØRMER (1942, pp. 152—157) used this principle in his studies of the phylogeny of the trilobites, and DE BEER (1940) indicated that it is genotypical changes of this kind which are responsible for the formation of higher taxonomic units (above genus).

In comparing the bifid and the trifid cardinal processes, there is also another problem which can be pointed out. In the Sowerbyellinae the lateral elements of the cardinal processes are regarded to be a reduced chilidium ("chilidial plates"). If this was true, these elements can not be analogous with the bifid cardinal processes. In all Strophomenidae sectioned by the author, the chilidium is histologically different from the rest of the valve (cf. SPJELDNÆS 1957, textfigs. 40—41), and this histological difference is not found in *Sowerbyella*. This indicates that the so called "chilidal plates" are analogous with the socket plates, and that the real chilidium is completely or almost completely reduced in *Sowerbyella*. This is also supported by the fact that a trifid cardinal process is found in at least one Leptestiniid with a chilidium (cf. *Ptychoglyptus valdari* SPJELDNÆS 1957, textfig. 17).

In this species there is no connection between the chilidium and the trifid cardinal process. Since it is highly unlikely that two different types of trifid cardinal processes occur in the Leptestiniids, it seems natural to conclude that the "chilidal plates" in the Sowerbyellinae (and some Leptestiinae) are analogous with the cardinal processes in the Strophomenidae, both being formed from the proximal ends of the socket plates.

In many highly developed Strophomenids, it is difficult to discriminate between the socket plates and the chilidium without a thin section because they are ancylosed. In the primitive species, such as *Rafinesquina*(?) *llandeiloensis*, the two different elements are clearly discernible (Pl. 1, figs. 2, 4, 5).

The question of the origin of the family Strophomenidae is still unsolved. *Rafinesquina*(?) *llandeiloensis* belongs to the *Kjaerina-Hedstroemina-Rafinesquina-Stropheodonta*-lineage (cf. SPJELDNÆS 1957, pp. 194—196, textfig. 42). In the other Strophomenid lineage, that of *Oepikina*, the Sowerbyellin cardinal process is still present, but the lateral elements are much larger than the medial one. This might be interpreted in two ways.

1) The retention of the passage of the pedicle to the ventral valve might have been less prominent in this lineage, leaving time for the development of a central cardinal process also. In this case the family

Strophomenidae might be monophyletic, since both lineages have the same feature (retention of the pedicle movement), even if it is of different degree in each of them.

2) The *Oepikina*-lineage of development of the cardinal process might also be interpreted as a simple development from the Sowerbyellin type, by more rapid growth of the lateral elements of the cardinal processes. In this case the two lineages are not at all related, and might have their origin in two different Plectambonitid groups.

In order to solve this problem, it is necessary to study a large number of larval valves of *Oepikina* and related forms. Until this is done, it is not possible to tell definitely if the family Strophomenidae is monophyletic, or if it is polyphyletic.

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EXPLANATION TO PLATE 1

Figs. 1—7. *Rafinesquina*(?) *llandeiloensis* (DAV.). All specimens from the zone of *Marrolithus inflatus incipiens* at Llan Mill, near Narberth, Carmarthenshire. *Fig. 1.* Ventral interior, 5x. *Figs. 2, 3, 5.* Larval dorsal valves. (Figs. 2, 5, 20x, fig. 3, 12x.) *Fig. 6.* Small, but well calcified ventral interior. Note absence of distinct muscle impressions, and marginal thickening. 3.5x. *Fig. 7.* Complete dorsal interior showing geniculation. 1.7x.

Figs. 8—9. *Rafinesquina* (?) *macallumi* REED. From the *confinis*-grits at Minuntion, Stinchar Valley, Girvan District, Scotland. *Fig. 8.* Fragmentary ventral interior. 3x, *Fig. 9.* Fragmentary dorsal interior. 3x.

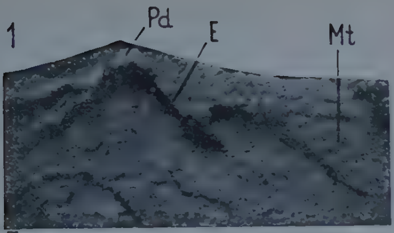
Fig. 10. *Oslomena* (?) sp. Cast of ventral interior, 1.7x. Loc. and hor. as in figs. 8—9.

All figs. except 10. are from latex impressions of casts. Figs. 1, 2, 3, 6, 7, 10 are whitened with ammonium chloride.
The specimens figured on figs. 1—7 belong to Geological Survey and Museum, London, no. 85415, the others belong to Paleontological Museum, Oslo.

Abbreviations: *Ch* = Chilidal plates, *Cp* = Cardinal processes, *Cr* = Pseudocrenulations on the socket plates. *E* = Elevated ridges of socket plates. *Mt* = Marginal thickening of the valves. *Pd* = Pseudodeltidium. *Sc* = Sockets. *Sl* = Slit between the cardinal processes. *Sp.* = Socket plates.

Plate 1

1



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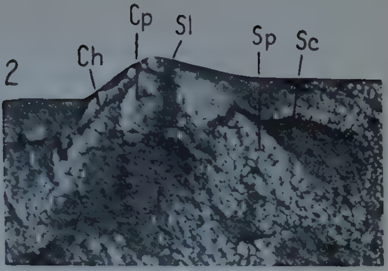

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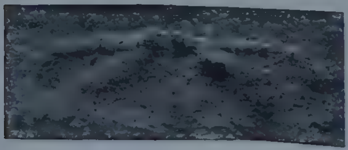

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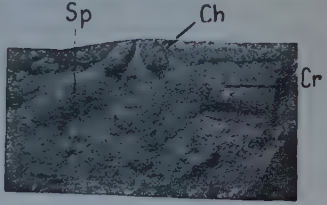

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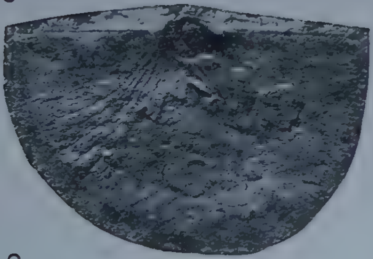

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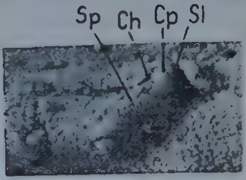

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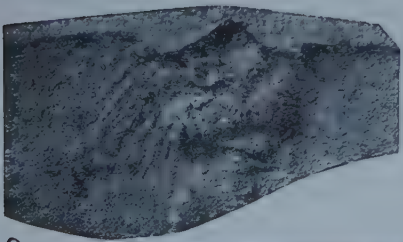

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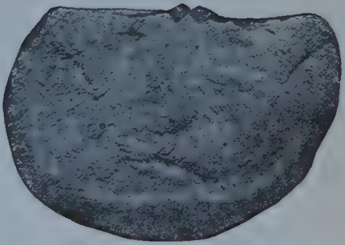

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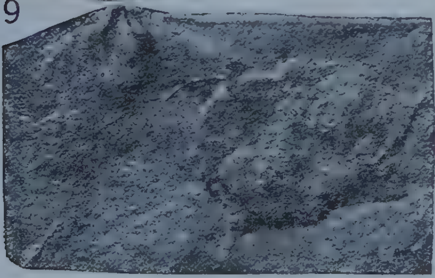

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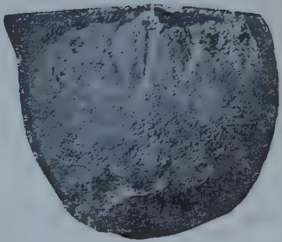

LA FLEXURE AXIALE DE LA DRIVA ET QUELQUES PROBLEMES STRUCTURAUX DES CALEDONIDES SCANDINAVES

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Avec 4 dessins par A. BAER

A la mémoire de C. W. CARSTENS

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Introduction.

Pour bien saisir la portée des problèmes exposés dans ce rapport, il est utile de les placer dans leur cadre. Celui-ci est plus large que la description régionale; il fait partie des principes de base des recherches tectoniques. C'est dans ce but que nous esquisserons d'abord brièvement ce cadre; nous essaierons ensuite de situer les recherches dans cet ensemble et nous en dégagerons quelques résultats en utilisant les principes d'interprétation exposés. Cette esquisse du cadre est nécessaire parce que les même objets prennent une signification différente suivant l'angle de vision. Cette signification influence, entre autres, aussi la terminologie, de sorte qu'une expression peut avoir un sens différent suivant le point de vue de l'auteur. Il est donc nécessaire de définir ce point de vue.

Pour arriver à des règles générales sur la genèse des chaînes de montagne, on a comparé des segments, soit du même âge, soit de divers âges.

On peut les comparer de plusieurs manières; ou bien d'une façon globale (les Alpes, les Pyrénées, les Appallaches, etc.) ou alors, on peut choisir différents points de vue: la nature de leurs matériaux, leur anatomie, leur cinématique, leur direction, leur évolution et d'autres aspects particuliers. La genèse et l'évolution sont naturellement les points les plus hypothétiques parce que les points à comparer ne sont pas des données immédiates, mais déjà des déductions à un degré supérieur.

Etudes comparatives des chaînes de montagnes.

Puisqu'une grande partie des énoncés de la tectonique est basée sur la comparaison des chaînes, il est nécessaire de toucher brièvement ce point. Les comparaisons se font de plusieurs manières. Certains auteurs traitent les chaînes de montagnes presque comme des individus dans le sens de la biologie, d'autres les considèrent dans le sens des espèces telles que la minéralogie les définit. Les auteurs ne définissent généralement pas leurs notions d'espèce d'une façon explicite; le fond de leur pensée apparaît dans la nomenclature et dans la manière de s'exprimer. Les notions générales restent le plus souvent subconscientes et ne sont pas analysées, parce que l'on admet que tout le monde les partage.

Il serait peut-être parfois bon de définir les unités que l'on se propose de comparer. Les ensembles de caractères distinctifs sont de nature très différente. Le groupe touchant de plus près la réalité observable est *l' anatomie actuelle* et ses rapports avec la morphologie; mais cet ensemble ne comporte que les renseignements d'un seul moment, le *moment actuel*, donc *une seule image dans la longue série qu'est l'évolution géologique.*

Certains auteurs semblent considérer ce moment comme l'aboutissement de l'évolution géologique, comme un état final; leur terminologie le suggère et l'exprime parfois.

Les chaînes de montagnes sont pour certains auteurs des zones *prédestinées* à être plissées pendant une certaine époque de l'histoire géologique (STILLE) et cette prédestination date de l'origine du globe.

Pour d'autres auteurs, ce sont les phénomènes d'un autre genre. Les limites des chaînes se déplacent dans le temps et dans l'espace de sorte que ce qui est considéré ailleurs comme unité, change d'une époque à l'autre. Les matériaux passent par des séries d'états, enregistrent ces passages et en conservent des traces à diverses échelles (WEGMANN, 1956) et de différentes manières.

L'état actuel (qui est un bon caractère distinctif) peut apparaître encore sous un autre angle. Notre époque est pour certains auteurs différente des états du passé, parce qu'ils ne la voient pas dans la perspective raccourcie de l'enregistrement géologique. En intégrant les faits de la tectonique vivante dans l'ensemble de l'évolution, on s'aperçoit que la structure des évènements, reste la même (WEGMANN, 1936, 1937), que le moment actuel n'est qu'une des images dans l'interminable série des états passés et futurs (HAWKES, 1958).

Des zones structurales s'ajoutent et se séparent d'une chaîne de montagnes au cours de son évolution, de sorte que *l'unité spatiale n'est pas conservée.* Elle n'existe que pour un moment défini, par exemple le présent. En tenant compte des hypothèses et principes énoncés et des possibilités non publiées, on s'éloigne vite des définitions simples assez répandues dans la littérature géologique.

Si on veut classer la notion des chaînes de montagnes au sens géologique, elle se place dans la catégorie des *séries d'états*, comme par exemple en biologie, les maladies. Cette série d'états s'étend à des espaces et des matériaux qui ne restent pas les mêmes; les états et les matériaux sont en mouvement, les uns par rapport aux autres, et ce sont ces mouvements qu'il s'agit de reconstruire.

La variabilité des conceptions ne se fait pas seulement sentir dans les exposés des résultats, mais aussi dans la manière d'observer et de sélectionner les phénomènes, et d'interpréter les observations et les mesures.

La nature flottante des concepts de base apparaît, entre autres, aussi par le fait que *deux courants* se dessinent dans la tectonique comparée: l'un d'entre eux ne cherche que les caractères communs aux diverses chaînes. Ce faisant, il tend à réduire toutes les chaînes au schéma d'un seul type ou de deux. C'est le point de vue uniformiste. L'autre tendance, analytique, essaie de souligner, à côté des ressemblances et des caractères communs, les traits individuels de chaque chaîne.

La tendance uniformiste apparaît par exemple dans les travaux de beaucoup d'auteurs ayant pris leur départ dans la chaîne alpine (SUESS, HAUG, ARGAND, KOBER, STAUB, KRAUS et d'autres). Ils essaient de ramener toutes les chaînes à leur image des Alpes, (qui varie d'ailleurs d'un cas à l'autre) en pressant les données dans la nomenclature et le schéma alpins. En sélectionnant ainsi les traits communs, on peut les rassembler en un ou deux types. Ces types ont rendu de grands services à la recherche, mais ils ont aussi caché des traits importants, car les types de ce genre ne représentent que l'extrait des observations sélectionnées et rassemblées en une image.

De l'autre côté, la comparaison des traits communs, des traits individuels et de leurs rapports, permet de dégager une réalité plus complexe. Vues sous cet angle, les chaînes ne sont plus que partiellement des unités semblables. Elles correspondent à des phénomènes et des ensembles de symptômes qui doivent être classés par des tableaux présentant un *grand nombre d'entrées.*

Les unités de comparaison ne sont plus les chaînes prises in globo, mais ce sont les aspects partiels que l'on rencontre à un stade beaucoup plus poussé de l'analyse. La classification des ensembles se fait par la combinaison des données élémentaires qui les composent. Les rapports entre les données élémentaires fournissent d'autres bases de comparaison. Celle-ci se fait donc, pour ainsi dire, d'abord à l'échelon des éléments de la mosaïque et de leurs rapports, avant de comparer les grandes lignes qui semblent se présenter d'une façon différente pour chaque génération. Dans notre cas la mosaïque ne se compose pas seulement d'éléments du même ordre de grandeur, mais chaque élément est à son tour composé, à plusieurs reprises, d'unités plus petites. L'image devient de ce fait moins facile à saisir. Elle échappe aux simplificateurs, puisque son caractère le plus apparent est sa différenciation.

Il y a eu de nombreuses discussions pour savoir si l'une ou l'autre des tendances mentionnées était préférable. Puisqu'elles fournissent des images d'une structure et par là d'une utilité fondamentalement différente, on doit se décider dans chaque cas, pour savoir la voie que l'on veut suivre. On se laissera guider par le *but* que l'on se propose. La simplification se prête mieux aux exposés didactiques; la manière analytique permet de trouver de nouveaux rapports entre les phénomènes.

Il ne faut pas se faire d'illusions sur la validité des déductions basées sur les types uniformistes. Le nombre des chaînes dont la structure et l'évolution sont bien connues n'est pas grand. Les conclusions basées sur un petit nombre et sur une sélection qui n'est peut-être pas du tout représentative, sont nécessairement peu certaines. Les règles que l'on peut en déduire, ont donc une structure logique tout à fait différente de celle des lois de la physique, parce que: 1) elles se basent sur un petit nombre de cas sélectionnés d'une façon discutable, et 2) elles sont déduites de cas qui représentent des ensembles de faits très complexes. Ce sont les résultats d'interférences compliquées. Seul le fonctionnement d'un petit nombre de facteurs a pu être isolé; la majorité des facteurs et des interférences n'est pas encore analysée.

Les comparaisons précédées d'une analyse de chaque cas, sont le seul moyen de circonscrire l'importance des facteurs et de les contrôler éventuellement par la voie expérimentale.

On peut procéder à des comparaisons dans différents plans. Chacun d'entre eux fournit des renseignements d'un autre genre. On peut comparer les chaînes: a) sur le plan des *inventaires* des matériaux et des phénomènes; b) sur le plan du *rapport entre les inventaires et les interprétations;* la connaissance des principes d'interprétation est naturellement une condition préalable; on est étonné de constater à combien d'interprétations différentes les mêmes phénomènes ont donné lieu; la terminologie en est souvent le reflet; c) sur le plan de l'évolution des recherches et connaissances, ce qui revient à connaître *la succession et la coordination des méthodes d'investigation et des principes d'interprétation* appliqués à un pays déterminé. Cette succession varie beaucoup d'une chaîne à l'autre. Elle détermine, d'une façon décisive, l'image que l'on peut actuellement se faire d'un segment de déformation. Il n'est pas possible de comprendre une telle évolution sans la situer par rapport à l'histoire de la géologie. La ligne de l'évolution varie d'une région à l'autre. C'est elle qui permet d'expliquer beaucoup de différences dans les conceptions de la structure des chaînes. Ces différences ne sont donc pas des différences réelles; elles ne s'expliquent que par l'état d'évolution des connaissances. C'est par des positions de départ très différentes que l'on a commencé à explorer et à dessiner l'image des différentes chaînes. Sans une analyse approfondie, il est difficile de *distinguer les différences réelles des chaînes des différences de perspective.*

Cette situation complexe correspond aux deux aspects principaux de la géologie: la perspective historique d'une part, et l'enchaînement des règles et des lois de l'autre. Comme connaissance historique, la géologie essaie de saisir l'évolution individuelle; elle met les événements en séries et elle étudie leurs rapports. Comme science naturelle, elle essaie de dégager les traits généraux et d'en tirer des règles et des lois. L'interaction des deux tendances déterminera pour chaque région et pour chaque ensemble de problèmes l'enchaînement des découvertes et le cheminement de la pensée synthétique.

Afin de donner une idée de l'évolution des connaissances dans divers tectogènes, nous en choisirons parmi les nombreux groupes de types, un pour les chaînes récentes et un pour les segments précambriens. Nous exposerons brièvement la suite des opérations telle qu'elle s'est présentée.

Deux types de succession des méthodes.

Les chaînes récentes montrent généralement des couches fossilifères permettant des études stratigraphiques. La succession des couches et les changements des matériaux sédimentés sont le point de départ pour les investigations structurales (WEGMANN, 1955, p. 188). Le Jura et les Alpes Occidentales représentent des exemples typiques. Les surfaces stratigraphiques permettent de construire une première image géométrique. L'analyse structurale développa par la suite ses propres méthodes. L'étude des grandes formes structurales mises en oeuvre par SCHARDT, LUGEON et ARGAND a ouvert de nouvelles possibilités et a permis de s'attaquer à des ensembles métamorphiques dépourvus de fossiles. Dans certaines régions des formes structurales ont été ainsi découvertes longtemps avant que leur explication ne fut possible.

L'analyse des formes existantes a été souvent confondue (de la part de certains auteurs allemands et autrichiens) avec l'interprétation génétique, ce qui a donné lieu à des discussions épiques. Certains auteurs n'ont pas hésité à affirmer que seules devaient exister des formes dont la genèse pouvait être expliquée par un procédé physique simple. Les possibilités d'explication sont ainsi limitées. La variabilité des formes est rationnée de cette façon et adaptée à l'étendue des connaissances de ces auteurs. Il est probable que ce soit là leur but, conscient ou inconscient. Ceci n'empêche pas la nature de nous mettre en face d'une infinité de variations. Le fait que l'on en découvre chaque année de nouvelles devrait nous rendre conscients qu'il existe une riche réalité en dehors des limites de nos schémas et qu'elle ne se révèle qu'à ceux qui en élargissent sans cesse les cadres. Des schémas restrictifs de ce genre se retrouvent encore actuellement dans la littérature. Leurs auteurs semblent oublier que les formes tectoniques ne représentent en général pas seulement l'enregistrement d'un seul événement, mais d'une multitude de phases dont les traces se superposent. Chaque événement d'une série de ce genre est à son tour le résultat d'une interférence d'un grand nombre de facteurs. *L'évolution d'une forme structurale ne correspond pas à une expérience physique, mais à une grande série d'expériences dont les effets se sont superposés.* Leurs traces ne sont que partiellement conservées. Une grande partie est effacée. On a mis au point un certain nombre de méthodes visant à reconstituer les parties manquantes; actuellement cette partie de la tectonique a besoin d'être développée.

Il s'agit dans ce cas tout d'abord de démêler l'enchevêtrement dans l'espace et dans le temps. Comme dans un noeud compliqué, il y a des trames-clefs qui doivent tout d'abord être défaites. Si on réussit à les détecter et à les résoudre, on pourra essayer, avec une certaine chance de succès, de défaire l'ensemble de la tectonique.

L'étude du métamorphisme ajoute une nouvelle perspective à celle de la stratigraphie et de la tectonique. Elle permet de fixer, dans beaucoup de cas, les rapports entre la déformation, les transformations et les migrations. Une nouvelle grille chronologique basée sur les données pétrographiques peut être établie de cette façon. Les phénomènes de l'ultramétamorphisme (SEDERHOLM), du plutonisme et du volcanisme, ajoutent de nouvelles possibilités de subdivisions chronologiques. Puisque ces phénomènes évoluent aussi dans l'espace, il est nécessaire de les rattacher aux formes géométriques des autres séries de données.

Chaque série nous fournit, pour ainsi dire, d'une part, une *grille chronologique*, d'autre part, *une grille spatiale*. Il est nécessaire de mettre en rapport les deux séries. Des méthodes de coordination ont été développées, mais il reste encore beaucoup à faire dans ce domaine. La littérature géologique montre beaucoup d'exemples de coordination très primitifs et il semble parfois que ceux-ci sont admirés par un plus grand public que les procédés plus nuancés et d'un emploi plus délicat.

La connaissance des chaînes précambriennes, sans chronologie basée sur les fossiles, a commencé en un temps où ces terrains étaient considérés comme la croûte primitive de la terre, d'abord par les neptunistes, ensuite, par les plutonistes. Elle a débuté par des déterminations minéralogiques et pétrographiques. Les images des formes structurales dépendaient entièrement des idées que l'on se faisait sur la mise en place des roches cristallines. Elles étaient des déductions théoriques sans contacts avec l'analyse géométrique et cinématique.

Les terrains cristallins furent considérés par beaucoup de pétrographes et de géologues (en partie encore actuellement) comme le domaine exclusif de la pétrographie. De nouvelles perspectives furent ouvertes par les études de J. J. SEDERHOLM. Son but essentiel était l'étude de l'évolution géologique des terrains précambriens, et en première ligne, leurs rapports chronologiques. L'introduction du principe de l'actualisme permit de reconnaître sous un camouflage métamorphique la nature primitive des formations sédimentaires et éruptives et de dévoiler les grandes lignes de la stratigraphie. Dans beaucoup de cas il était possible de séparer les épisodes de l'histoire supracrustale de ceux de l'évolution intracrustale.

La succession des changements intracrustaux obtenue par les études pétrographiques reste le plus souvent sans données spatiales. Elles se limitent à déterminer les changements de phases, de compositions chimiques et minéralogiques. Elles essaient de déduire de ces données les températures et les pressions ayant présidé aux transformations.

La détermination des températures et des pressions est d'un intérêt limité, s'il n'est pas possible de relier ces données à une image spatiale et cinématique. Il fallait donc introduire la notion de l'espace à différentes échelles (de l'affleurement, de l'unité tectonique et des segments de déformation). Ce fut le rôle d'un groupe de méthodes et techniques, tout d'abord inspirées par la tectonique alpine (WEGMANN, 1929) mais adaptées par la suite aux conditions des étages des socles cristallins. Dans chacun des étages les matériaux enregistrent les conditions qui y règnent d'une manière différente. Les matériaux passant par plusieurs étages sont caractérisés par la *superposition des traces enregistrées* pendant leur périple. La succession de ces traces ne peut pas seulement être suivie dans le temps, mais on peut aussi lui donner des coordonnées approximatives dans l'espace.

L'évolution des connaissances tectoniques dans les socles cristallins a donc cheminé d'une façon différente de celle des terrains fossilifères Dans ces derniers elle prend son départ en s'appuyant sur des surfaces de sédimentation, généralement immédiatement observables, tandis que les témoignages de l'épisode intracrustal doivent être dégagés en partant des traces qu'ils ont laissées. Dans les socles cristallins ce sont les traces des phases *intracrustales* qui sont apparentes, tandis que l'évolution supracrustale doit être déduite en dégageant l'image de la stratigraphie sous le camouflage du métamorphisme et de l'ultramétamorphisme. *L'entrée en matière et la succession des recherches sont donc très différentes dans les chaînes récentes et dans les vieux socles.* Si on veut comparer les données fournies par les deux groupes de terrains on doit tenir compte de cette différence d'accès aux sources de renseignements. Cette différence d'origine confère à chaque groupe sa propre structure logique. Il est clair que la tectonique des socles est plus difficile à reconnaître et qu'elle exige une connaissance plus approfondie des méthodes et des techniques. Une première condition est la *connaissance des types structuraux des terrains non-métamorphiques*

Nous avons réduit les deux types d'évolution à leurs lignes les plus simples. Il est maintenant possible de placer dans ce cadre l'image particulière des investigations de la chaîne calédonienne scandinave et plus particulièrement celle de la région considérée.

Evolution des recherches dans la chaîne Calédonienne Scandinave.

Cette évolution est très intéressante et instructive. Une série de notions fondamentales et de nombreux noms célèbres s'y rattachent. Le cheminement de la pensée est peu connu en dehors de la Scandinavie, mais certainement très intéressant au point de vue général, car sa portée dépasse le cadre régional. Les épisodes d'investigations pétrographiques, stratigraphiques et structurales se suivent dans la chaîne calédonienne d'une façon assez différente de celle des chaînes récentes et des segments précambriens. Elles ont pris un chemin un peu différent du côté norvégien et du côté suédois. Nous essaierons d'indiquer quelques problèmes sur lesquels l'attention des géologues scandinaves s'est dirigée tour à tour et pour lesquels ils ont trouvé des solutions d'un intérêt général.

Sans énumérer toutes les découvertes importantes, nous nous contenterons de mentionner quelques groupes: Un premier est surtout caractérisé par les noms de KJERULF, de KEILHAU et de DAHLL. C'est le *premier inventaire des faits et leur classification.* Celle-ci est déterminée en partie par les idées de l'époque, en partie par les conditions spéciales de la région qui ont frappé les auteurs. Notons-en un exemple: les observations et les connaissances physico-chimiques ont amené KEILHAU en 1836 à admettre qu'une partie des roches granitiques et gneissiques était formée par transformation à partir de roches préexistantes sans l'intermédiaire d'une phase liquide importante (citations, cf. WEGMANN, 1948, p. 46).

Le prochain stade est caractérisé par les *grandes synthèses tectoniques* de TÖRNEBOHM (1896) et les *études structurales* de REUSCH (1888). Elles ont abouti à une image d'ensemble de la tectonique; celle-ci, basée sur les connaissances de l'époque, se trouvait à l'avant-garde du progrès. Elle est imposante par la hardiesse des conceptions géométriques. La superposition de couches hautement métamorphiques sur les sédiments fossilifères marginaux, constatée dans la province suédoise du Jämtland et les districts voisins de la Norvège, suscita l'hypothèse des grands charriages. L'essai de grouper l'ensemble des observations en une image géométrique consistante est caractéristique pour ce stade.

Il fut suivi par une des plus brillantes périodes de l'évolution de la *pétrographie.* Celle-ci fut déclenchée par *l'application des données de la chimie physique* et inaugurée par W. C. BRØGGER et J. H. L. VOGT. Cette pétrographie basée sur la règle des phases, concentra son attention sur les réactions accompagnant la transition de l'état de fusion à l'état cristallin. *L'état fondu devenait ainsi le point de départ des phénomènes explicables* et il est naturel que l'on en chercha partout la présence. On s'intéressa surtout aux roches intrusives dont on pouvait admettre qu'elles étaient mises en place à l'état fondu. Pour beaucoup de roches qui n'ont jamais été fondues, une origine liquide fut admise pour pouvoir appliquer les règles et les lois trouvées. Le lieu d'origine de ces roches se trouvait dans les profondeurs inconnues de l'écorce terrestre. Les mouvements de ces masses fondues ne semblaient pas obéir à des règles cinématiques bien définies, à l'exception des "pluies de cristaux dans les réservoirs magmatiques». En tout cas on ne s'en inquiétait pas. Les observations concernant la solidification du liquide furent sélectionnées et leur explication fut cherchée dans le domaine physico-chimique, ce qui déviait, pour ainsi dire, l'attention des ensembles géométriques et de leur configuration dans l'espace. Les découvertes dans le domaine de la pétrographie, de la minéralogie et de la géochimie, ont porté le nom de l'école d'Oslo dans le monde entier, surtout par les travaux de V. M. GOLDSCHMIDT, de ses collaborateurs et de ses successeurs.

Les écoles pétrographiques, aussi bien en Norvège qu'en Suède, abandonnèrent les idées directrices de TÖRNEBOHM et de REUSCH. En réduisant autant que possible les charriages, elles expliquèrent la position des terrains hautement métamorphiques sur les couches fossilifères, par des intrusions et leur influence métamorphisante. Celle-ci devait se faire sentir dans les parties supérieures envahies, en laissant intactes les couches basales. Puisqu'il fallait réduire les charriages, on était obligé de fermer aussi vite que possible les charnières, souvent directement derrière l'affleurement. *Les directions de transport* devenaient de ce fait *extrêmement variables;* les mouvements étaient dirigés dans toutes les directions du compas et cela même pour des régions très limitées. Le sens des transports était toujours (dans cette théorie) du bas en haut sur l'inclinaison des couches, de la foliation ou de la schistosité. Ceci revenait à dire que *les axes étaient sensiblement horizontaux* (cf. les coupes de cette période p. ex. FRÖDIN, 1922). La possibilité des axes inclinés fut considérée, à ce temps-là, comme une complication contraire à l'économie des principes de base.

Arrivé en Norvège en 1924, grâce à une aimable invitation de C. W. CARSTENS, nous fûmes, depuis le commencement, frappés par l'imposante tombée axiale le long de la vallée de la Driva. Ce phénomène tectonique fut étudié à plusieurs reprises en 1924 et 1925 entre la région de Dovre et le fjord de Trondheim. Les régions homologues, plus au Nord et plus au Sud, furent visitées pendant ces années. Ces observations permirent d'esquisser les grands traits de la *position spatiale des axes* (WEGMANN, 1925, 1926).

Un certain nombre de *culminations et de dépressions axiales* furent reconnues. Sur ces oscillations de premier ordre, se greffent des ondulations d'importance secondaire. La connaissance des régions à inclinaison axiale est nécessaire pour construire des coupes correspondant à une certaine profondeur. Le principe des constructions fut exposé par ARGAND (1911) pour les Alpes penniques. Même si toutes les unités rencontrées dans une coupe de ce genre ne continuent pas tout au long du segment déformé, les régions d'inclinaison axiale peuvent toutefois livrer une *formule structurale.* Les variantes de cette formule peuvent, dans certains cas, être déterminées ou conjecturées par d'autres observations. La connaissance du *style tectonique* est particulièrement importante. On peut définir ce style comme le rapport des formes des grandes unités tectoniques avec celles des unités plus petites en passant par les échelles de l'affleurement, de l'échantillon, jusqu'aux images microscopiques. Cette notion forme la base rationnelle d'une subdivision chronologique des mouvements intracrustaux et par là, de la reconstitution cinématique (WEGMANN, 1951). Elle permet aussi, dans beaucoup de cas, de distinguer les étapes tectoniques, leur superposition et leur inversion.

Parmi les montées axiales calédoniennes, celles des *Lierne—Namsos* et de la *Driva* sont particulièrement intéressantes. Elles encadrent la *dépression axiale de Tröndelagen et du Jämtland,* et permettent de saisir la *virgation* correspondant à ce secteur de la chaîne calédonienne. Ce sont donc deux régions-clefs. Elles furent étudiées dans les années 1924 et 1925. La conception structurale était trop différente des préoccupations pétrographiques de ce temps, de sorte que les descriptions sont restées dans les tiroirs et les cartons. Seuls des résumés purent être publiés (WEGMANN, 1925, 1926).

Les idées tectoniques ayant évolué entre temps en Scandinavie, les résultats des anciennes recherches commencèrent à gagner une certaine actualité. Avant de les réaffirmer, nous avons tenu à les vérifier par quelques contrôles dans la région-type de la Driva.

Grâce à l'amabilité et au talent d'organisation de MM. TOM. F. W. BARTH et JOHANNES A. DONS, nous avons pu, accompagnés par des élèves de Suisse et de France, passer une semaine dans cette région et y obtenir un grand nombre d'observations et de mesures. M. DONS a bien voulu nous accompagner et nous rendre le séjour aussi agréable et intéressant que possible. Il a bien voulu nous encourager à rédiger ce rapport sur les problèmes, les observations et quelques conclusions qui semblent s'en dégager. Ces conclusions sont provisoires, mais elles sont de nature à attirer l'attention sur quelques faits, restés jusqu'ici dans l'ombre, faits qui peuvent avoir une répercussion dans plusieurs domaines, plus particulièrement dans ceux de la géologie structurale, de la stratigraphie, de la paléogéographie et de la géochimie. Nous remercions vivement MM. Barth et Dons de toute leur gentillesse et de leur intérêt. M. Per Holmsen a bien voulu nous sacrifier une journée et nous fournir des renseignements précieux; qu'il veuille trouver ici l'expression de notre vive reconnaissance.

Nous exprimons aussi nos remerciements à tous les participants du camp d'exercices tectoniques de 1957 à Kongsvoll, tout particulièrement à M. A. Baer qui a bien voulu dessiner les figures 1 à 3, pl. 1 et à M. J. P. Schaer qui s'est aimablement chargé de revoir le manuscrit et les épreuves.

Structure de la région.

On distingue depuis longtemps les groupes suivants de formations:

1) les gneiss de base,
2) les helleskiffer contenant des gneiss oeillés, des anorthosites et d'autres roches.
3) le groupe de Röros comprenant des micaschistes et des roches basiques,
4) le groupe des roches moins métamorphiques, appartenant à la série de Trondheim, forme les étages tectoniques supérieurs et se distingue par son style.

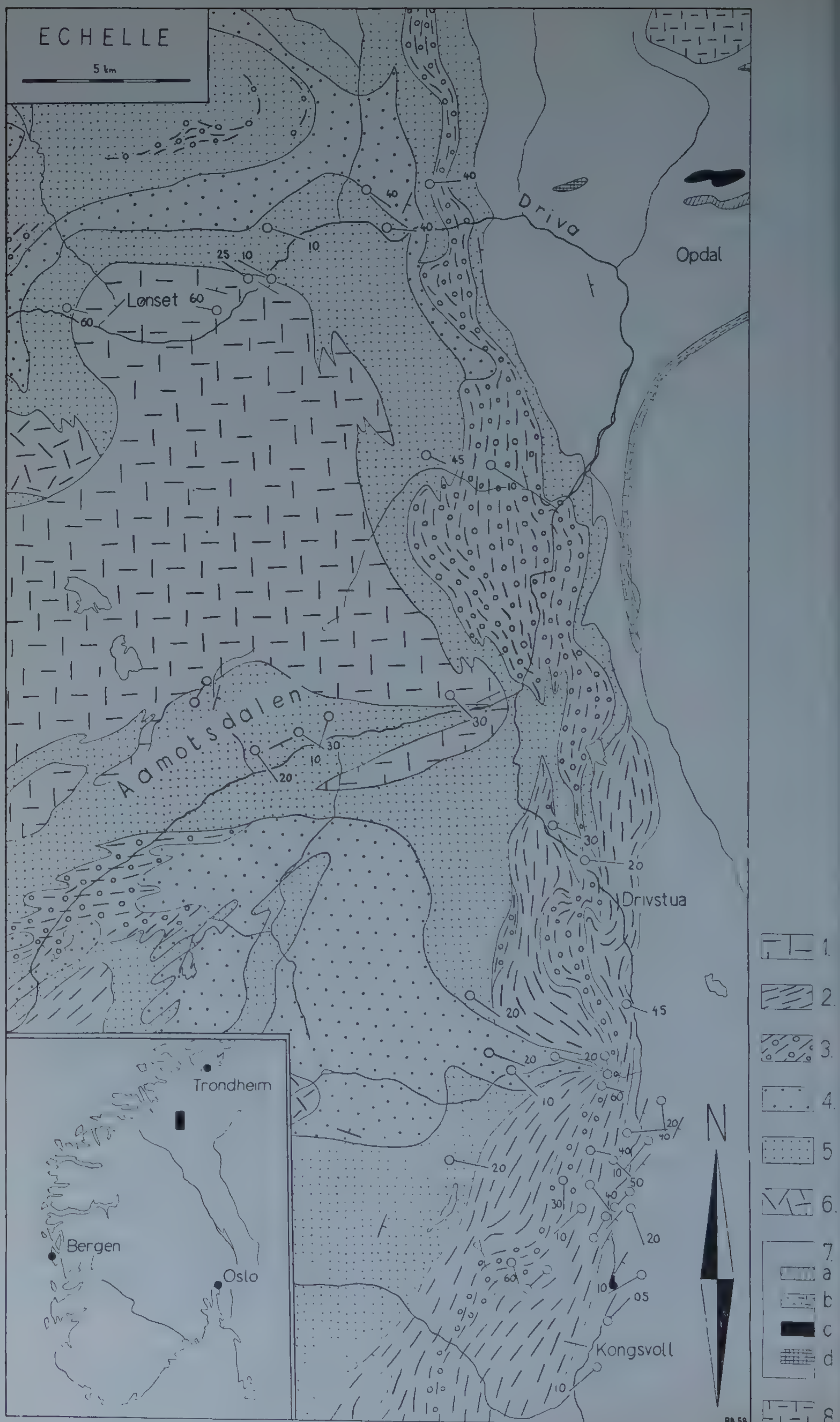
En ce qui concerne la stratigraphie et la pétrographie, nous renvoyons aux travaux publiés, dont un certain nombre est cité dans la bibliographie. Celle-ci n'a pas la prétention d'être complète.

La distribution des terrains est indiquée sur la carte fig. 1. Les indications des auteurs des travaux antérieurs et les résultats de nos observations y sont rassemblés (comp. fig. 2).

On y observe des formes de déformation à plusieurs échelles:

1) les grands plis dont le dessin est obtenu par la cartographie;
2) les plis à l'échelle de l'affleurement, qu'il est possible d'observer directement;
3) les plis à l'échelle de l'échantillon et l'orientation des éléments minéralogiques.

Les trois éléments montrent à première approximation des directions semblables. Les axes des formes qui dépassent l'échelle des


ECHELLE
5 km
Driva
Opdal
Lønset
Aamotsdalen
Drivstua
Kongsvoll
Trondheim
Bergen
Oslo
N
1.
2.
3.
4.
5.
6.
7.
a
b
c
d
8
AB 58

affleurements ont été obtenus à l'aide de la projection stéréographique. Des tendances à l'enroulement axial et aux structures hélicoïdales ont été constatées, mais nous n'avons pas eu le temps de les poursuivre.

L'orientation générale de ces axes correspond à une plongée vers l'ESE. En suivant les variations de plongement, on constate qu'à l'W (Aamotsdalen p. ex.), l'inclinaison est faible et régulièrement dirigée vers l'E. Dans le prolongement de cette direction, on voit brusquement plonger les axes (jusqu'à 40° et plus) en direction de la Driva. Au niveau du thalweg, se dessine par endroits une courbure inverse de la précédente, ramenant l'inclinaison axiale à 10°E seulement. La flexure axiale est donc importante. Elle se marque nettement dans la topographie, puisque le versant occidental de la vallée de la Driva est taillé à peu près parallèlement à la plongée axiale la plus forte.

Dans le détail, les mesures relevées dans le Stölådalen et le Kaldvelladalen sont particulièrement typiques. L'itinéraire suivi remonte la Stölå depuis son embouchure dans la Driva et rejoint dans l'ouest

←

Fig. 1. *Carte schématique de la région de Kongsvoll—Opdal* montrant la position des axes des plis mesurés.

Légende:

1) *Complexe antécalédonien*, dit «des gneiss de base» (granites, amphibolites et autres roches caractérisant le socle cristallin).

Complexe intermédiaire:

2) Gneiss indifférenciés.
3) Gneiss oeillés.
4) Schistes micacés.
5) «Helleskiffer», quartzites micacées.

Complexe de Trondheim:

7) Roches vertes, comprenant:
    a) quartzites jaspéroides
    b) calcaires
    c) «pillow-Lavas»
    d) kératophyre.

Roches mises en place après les phases de mouvement intime de leur cadre:

8) Opdalite.

NB. Stølådalen se trouve entre Åmotsdalen et Opdal, Nystugudalen entre Kongsvoll et Åmotsdalen.

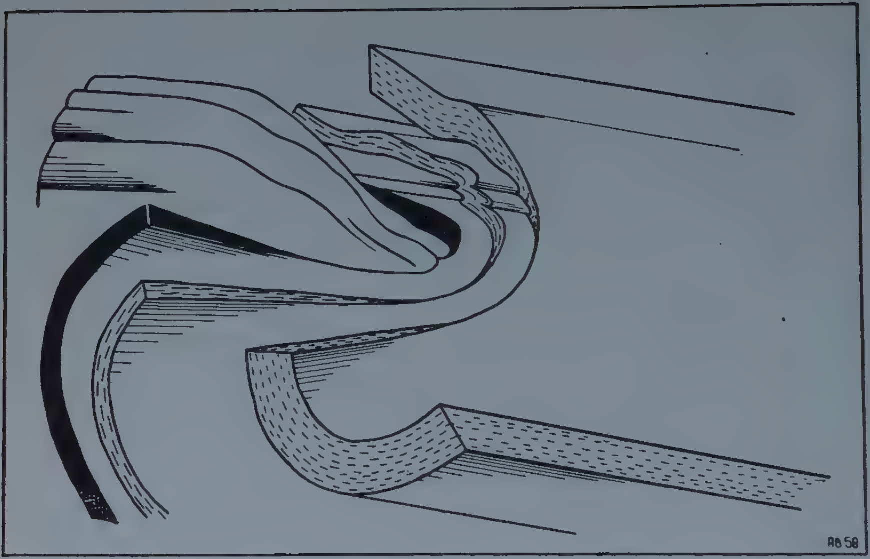


Fig. 2. *Stéréogramme schématique de la descente axiale du Drivdal.* Les gneiss de base plongent vers l'E (droite) sous leur couverture déversée au SSE. Les trois enveloppes du massif représentent schématiquement de haut en bas: le complexe de Trondheim, la zone de déformation des gneiss oeillés et (en noir) le complexe des «Helleskiffer» et des schistes micacés, avec leurs roches plutoniques. La puissance des couches dessinées ne prétend pas à l'exactitude, alors que les inclinaisons axiales correspondent à la réalité.

la partie supérieure du Kaldvelladalen. Les résultats ont été reportés dans le diagramme fig. 3. La flèche indique le déplacement des axes à mesure que l'on avance vers l'ouest. Au départ, à Stölån, ils sont N 120°E, inclinés de 10° vers le SE, puis ils plongent davantage sans changer beaucoup de direction et atteignent N 120°E, avec inclinaison 45°SE. Lorsqu'on s'enfonce plus à l'W encore, au N du Kaldvellsjö p. ex., le phénomène inverse se produit et l'inclinaison axiale retombe à 15°ESE (cf. fig. 3).

En plus des axes principaux, dont le comportement caractérise la région, on notera quelques axes subhorizontaux orientés vers le NNE (donc à peu près perpendiculaires aux premiers). Ils sont surtout marqués le long de la Driva. Ces plis secondaires, moins nets, accompagnent la brusque flexure des éléments tectoniques. Leur place dans la chronologie est marquée par le fait que les masses plongeantes sont parfois amincies et même faillées dans la même direction. Leurs plis sont plissés à plusieurs endroits. Le sens relatif du déplacement des masses rocheuses a pu être constaté à beaucoup d'endroits et par plusieurs sortes d'enregistrements. Les différents témoignages indiquent un transport des masses supérieures du NNW vers le SSE. Quant aux déformations, dont la trace s'étend sur la carte du NNE au SSW, elles correspondent à un soulèvement du bouclier situé à l'WSW par rapport à la dépression de Tröndelagen.

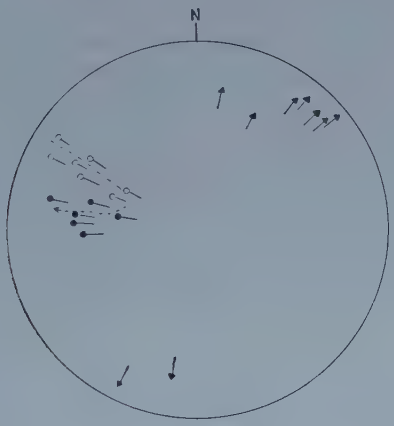


Fig. 3. Réseau stéréographique montrant les variations de l'inclinaison axiale dans la vallée de la Driva. Les axes à tête triangulaire sont les axes secondaires. Ils proviennent surtout du Nystugudalen. — Les axes à tête ronde vide sont relevés dans la partie inférieure du Stølådalen, ceux à tête pleine, au N du Kaldvellsjø. La flèche pointillée indique la succession des axes observés d'E en W en remontant le Stølådalen. Hémisphère sup.

Rapports chronologiques.

Les rapports de ces deux séries de mouvement sont assez clairs. Le soulèvement du bouclier est grosso-modo plus récent que le transport vers le SSE, bien que les deux épisodes se chevauchent en partie.

Le mouvement principal peut être subdivisé par diverses observations. Une marque chronologique importante est représentée par les filons basiques. Ils traversent la série des Helleskiffer. Après leur mise en place, ils furent fortement déformés comme on peut le constater dans la région des carrières entre Drivestuen et Rise. Ces filons ont été déjà observés par Björlykke (1905, p. 389, 1909). Nous les avons retrouvés en 1925 sur les rives du fjord de Trondheim où ils ont subi une déformation en partie moins intense que dans la vallée de la Driva. Ces filons ont fait intrusion dans une série déjà un peu déformée, mais avant le plissement final.

La série des Helleskiffer montre plusieurs formes de déformations:

a) la déformation interne des sédiments et des roches plutoniques qui les accompagnent. Les gneiss oeillés furent déjà connus par Leopold von Buch en 1810. C. W. Carstens (1925) montra qu'il s'agit en partie de granites déformés à structure de Rapakivi. Cette déformation se place pour une grande partie avant le métamorphisme décrit par Barth (1938), Rosenqvist (1944) et Holmsen (1955).

La tectonique à plus grande échelle permet d'autres subdivisions:

b) la série des Helleskiffer et des micaschistes contenant les roches plutoniques a glissé en formant des écailles superposées;

c) cette série d'écailles fut, à son tour, plissée en formant des plis dont la direction axiale est grosso-modo dirigée E-W;

d) cet ensemble a subi la flexure et un plissement dépendant de celle-ci, visible dans certains secteurs de la Driva et de sa continuation vers le nord;

e) le charriage de la série de Trondheim sur le groupe des Helleskiffer se place probablement entre c) et d). Il a subi les effets de d), mais surtout aussi ceux du groupe f);

f) sous le nom de groupe des mouvements tardifs sont compris des déplacements importants mais n'ayant pas laissé de trace ubiquiste de sorte que leurs empreintes peuvent sembler moins importantes étant confinées à des espaces restreints. Leur déplacement n'est pas de grande dimension. Ces mouvements montrent des caractères divers.

La distinction plus ou moins nette des stades permet de dérouler quelques épisodes de l'histoire de la chaîne calédonienne. Avant de tirer ces conclusions, il est nécessaire de mentionner plusieurs autres problèmes, dont celui du socle et celui des rapakivi. Notre but n'est pas de les résoudre, mais de les dégager des ornières et des culs-de-sac et d'ouvrir ainsi de nouvelles voies d'accès pour les études structurales.

Dans cette chronologie, nous n'avons pas suivi les vues des auteurs admettant que les gneiss oeillés du genre rapakivi et leur suite ont été mis en place pendant la déformation calédonienne. Nous pensons comme Törnebohm (1896) et C. W. Carstens (1925) que leur structure actuelle est due à la déformation de roches mises en place avant les mouvements calédoniens proprement dits.

On peut naturellement discuter sur ce que l'on veut désigner comme mouvements calédoniens. On peut rassembler dans ce groupe toutes les déformations commençant par celles qui ont affecté la surface du vieux socle lors de son établissement. On peut aussi limiter ce terme aux mouvements ayant laissé des traces dans le Cambro-Ordovicien. La seconde manière nous semble préférable pour des raisons de commodité.

Dans ce cas, la question suivante se pose: comment différencier les mouvements calédoniens dans le socle? Car si les formations Cambro-Siluriennes ont subi des déformations intenses, il est probable que le socle n'est pas resté épargné; il semble au contraire à première vue que le socle a subi un nombre plus grand de déformations. Il a été déplacé à plusieurs reprises d'un étage tectonique à un autre, et il a chaque fois subi des mobilisations et des recristallisations. Celles-ci sont de différentes espèces. Il n'est pas possible d'exposer ici toutes les observations faites en son temps et les réflexions qu'elles suscitent. Nous nous contenterons d'un résumé.

Remarque sur le rôle tectonique du socle cristallin.

Le rôle tectonique des socles cristallins a été interprété d'une façon très différente. L'évolution de la nomenclature peut en donner une idée. On a tour à tour considéré les régions gneissiques de l'Ouest de la Norvège et de la région de Namsos: a) comme un vieux bloc non influencé par les mouvements calédoniens ou ayant formé un massif résistant; b) comme étant entièrement, ou en tout cas en partie formé de roches intrusives d'âge calédonien.

Nous avons trouvé dans les années 1924 à 1926 une série de bandes de roches calédoniennes pincées dans les vieux gneiss (p. ex. péninsule de Fosen). Ces parties de roches sédimentaires et volcaniques sont encadrées par des zones de mouvements certainement calédoniens; d'autre part, on constate la présence de structures anciennes reprises par des dislocations plus récentes et d'autres témoignages de ce genre. Ces observations furent confirmées par la suite par d'autres auteurs (Strand, Gjelsvik, Hernes, Gleditsch et d'autres).

Actuellement il est difficile de considérer le cristallin du bouclier occidental et de Namsos comme des îles du socle Précambrien entourées de segments déformés d'âge calédonien. Le socle a subi plusieurs séries de déformations intenses. Il a joué un rôle important dans la tectogenèse calédonienne. L'étude des déformations et leur analyse chronologique permet de se faire une idée de ce qui se passe dans cet étage tectonique. Les études structurales ne couvrent actuellement qu'une petite partie de ces territoires de sorte qu'il n'est pas possible d'en donner une vue d'ensemble. On doit se contenter pour le moment de données partielles.

On peut se demander par quels critères il est possible de séparer les structures anciennes des déformations plus récentes. Il y en a toute une série, mais ils n'ont pas le même poids.

Une première donnée, une des plus importantes, est l'identification de la déformation de la surface d'érosion recoupant le socle et recouverte en transgression par des formations de couverture. Ces formations n'ont pas partout le même âge. Dans certaines zones, ce sont les schistes du Cambrien (schistes alunifères), dans d'autres, ce sont différents membres de la série Eocambrienne des sparagmites, ou encore éventuellement des formations plus anciennes. *La surface n'a donc pas partout le même âge.* Elle se compose de *facettes métachrones.* Il est probable qu'elle correspond, pour des régions limitées, à une pénéplaine. Il faut donc considérer cette surface avec les formations de couverture correspondantes. Ceci n'est pas partout possible parce que la couverture a été décollée et d'autres formations se sont mises à sa place.

Les formes actuelles de la surface d'érosion ancienne avec ses grandes dénivellations et ses surplombs portent les marques de la déformation et peuvent donner une première approximation.

Ces formes peuvent être reliées aux déformations internes. La structure semble peut-être au premier abord assez chaotique, par exemple dans la péninsule de Fosen entre le fjord de Trondheim et l'océan. Le raccord des renseignements obtenus par différentes méthodes permet parfois de faire la part de chaque groupe chronologique.

En circulant le long de la ligne du chemin de fer ou le long de la route de la vallée de Rauma, on constate une telle profusion de traces tectoniques superposées et une telle richesse de combinaisons que l'on croit se trouver dans un labyrinthe. Toutefois, en cherchant, on tombe parfois sur des endroits favorables permettant de défaire quelque trame de cet enchevêtrement. Ce sont les affleurements-clefs; c'est à partir d'eux qu'il s'agirait d'étendre les études. Cette région promet beaucoup de résultats d'ordre régional et général (tout particulièrement sur les étages tectoniques) et elle procurera des joies intenses au chercheur patient.

Les déformations internes des roches sont le plus souvent fixées par le métamorphisme, le même qui a transformé les formations de couverture pincées. Dans certaines régions, le socle a subi un ultra-métamorphisme suivant l'expression de SEDERHOLM (cf. STRAND, 1949, 1951). Ailleurs, les anciennes mylonites et autres roches déformées ont été recristallisées et parfois mobilisées. Ces zones ont actuellement l'aspect de filons et de zones intrusives.

Quand on parle d'un socle cristallin, on pense le plus souvent à sa manière d'être sous un pays tabulaire ou sous une série de couverture décollée. Le Jura et les Alpes helvétiques sont des exemples typiques de ce genre. Le style des séries décollées est dans ce cas plus souple que celui du soubassement cristallin. Ce dernier montre des formes caractérisant le «style cassant», comme par exemple les coins de la Jungfrau. Une telle configuration est caractéristique pour les étages tectoniques supérieurs et permet de les caractériser. Elle est réalisée pour la zone calédonienne dans la région du Mjösen (SKJESETH, 1953, 1954).

Si la déformation a lieu à une plus grande profondeur, les choses se passent autrement: les gneiss étant des roches feuilletées et polyminérales deviennent plus mobiles que les quartzites, certains conglomérats et autres roches de ce genre. Sous une série quartzitique il y aura donc de nouveau des roches plus mobiles. Elles se conforment à la tectonique de la série du toit et auront l'air *d'épouser le relief inférieur des quartzites*. Des phénomènes de ce genre sont connus dans beaucoup de chaînes depuis les Alpes jusqu'aux segments précambriens. Dans les Alpes Penniques, par exemple, la structure des quartzites du Trias est épousée par la formation des schistes lustrés (Jurassique) d'une part, et par celle des schistes de Casanna (Paléozoïque) de l'autre.

Si le phénomène se développe, les roches mobiles du bas peuvent s'introduire entre les autres et présenter des configurations caractéristiques. Le terme final représenté par le diapirisme (WEGMANN, 1929) est réalisé quand les roches deviennent très mobiles. Des exemples de ces différents stades sont visibles aussi bien dans la région des gneiss de l'W, que dans le massif de Namsos. La structure en pelure d'oignon de ce massif avec ses feuillets gneissiques intercalés de marbres, de micaschistes et de roches à disthène est caractéristique pour cette région située au sud du fjord d'Indre Folla. La remobilisation à plusieurs reprises est particulièrement intéressante dans les affleurements des formations profondes.

Les massifs limités en forme de coupoles arrondies ou allongées passant à des plis déversés de forme simple caractérisent cet étage tectonique. Ils se distinguent des plis simples des terrains sédimentaires par le fait qu'ils ne sont pas déroulables. Leur mécanisme cinématique est donc différent. L'anticlinal de Lønseth (ROSENQVIST, 1941, HOLTEDAHL, 1938, 1948, HOLMSEN, 1955) et les autres intumescences au NW et à l'W sont caractéristiques pour ce style. On le retrouve dans la région de Namsos - Indre Folla en Norvège et au nord du lac Ladoga en Finlande (TRÜSTEDT, 1907). Ces formes sont caractéristiques pour cet étage tectonique. Elles constituent pour ainsi dire un enregistrement typique du séjour des matériaux sous ces conditions. Elles sont parfois conservées en montant ou en descendant dans un autre étage. Des exemples où ces structures sont descendues ont été décrits par BERTHELSEN (1957) au Grönland occidental. Les matériaux ont changé de caractère étant «granulitisés» (au sens scandinave du terme).

Les coupoles et bourrelets font partie d'une série évolutive. L'arrêt à tel ou tel stade de cette série est caractéristique et montre bien qu'en général les conditions d'équilibre ne sont pas atteintes avant qu'une nouvelle situation s'amorce. Il en est du reste de même dans beaucoup de cas pour la pétrographie. C'est ce qui rend l'étude des structures superposées particulièrement attrayante (WEGMANN, 1947) (cf. fig. 4).

L'intérêt de ces structures ne peut être saisi que si l'on connaît: 1) leur *situation géométrique* dans l'ensemble, (ce qui n'est pas possible sans l'étude des culminations axiales); 2) leur *position dans l'histoire* qui peut être déchiffrée par l'analyse cinématique. Jusqu'à maintenant, on ne peut, pour ainsi dire, déterminer que quelques points d'une courbe certainement très compliquée. L'image que nous pouvons nous en faire est donc bien simplifiée. Citons quelques points importants:

Les stades antérieurs à la chaîne qui fut pénéplainée avant le cycle calédonien ne peuvent être reconstruits que d'une manière imparfaite. Il est possible, même probable que le socle se compose de plusieurs chaînes de divers âges. Il est en outre probable qu'il y ait, comme dans la plupart des chaînes, un étage intermédiaire entre les formations du socle proprement dit et les séries de la couverture sédimentaire. Nous reviendrons à ce problème.

A la fin de la première partie de l'histoire du socle l'érosion a mis à nu des étages autrefois profonds des anciennes chaînes. Les terrains qui avaient réagi auparavant d'une façon souple, sont ainsi arrivés dans un étage tectonique où ils réagissent avec un style cassant, ce qui se traduit dans les formes et les caractères des produits (mylonites, brèches etc.). Une anisotropie d'un genre spécial a pris naissance de cette façon et détermine le découpement plus récent.

Les mouvements enregistrés ainsi par les roches correspondent en surface à la formation de bassins et de surélévations. Les parties hautes sont attaquées par l'érosion, les bassins sont remplis par les produits sédimentaires et volcaniques. Quand ils ont atteint une certaine épaisseur, l'ancienne surface est de nouveau ensevelie. Le remplissage des bassins peut être d'origine sédimentaire, volcanique ou tectonique. Des masses importantes glissent dans les dépressions. De tels glissements n'ont pas seulement lieu à la surface, mais aussi en profondeur.

Les remplissages sédimentaires, volcaniques et tectoniques ont ramené le socle des Calédonides de nouveau dans des étages tectoniques inférieurs, mais avec une structure changée. Cet enfoncement correspond à une *nouvelle période de métamorphisme, d'ultramétamorphisme et de mobilisation,* diffuse ou en masse (WEGMANN 1935, STRAND, 1951). La montée de la chaîne se fait d'une façon différenciée. Les culminations axiales montent plus vite que les dépressions. Des phases tardives déforment et recourbent les plans de charriage, en un style plastique dans les étages tectoniques profonds, en un style plus cassant dans les étages plus élevés. C'est ainsi que de nombreux plans de charriage sont plissés depuis les bords de l'océan jusque vers Vaage et les bords du Mjösa. Dans cette région les enroulements se font d'après un style correspondant aux étages supérieurs (cf. SKJESETH, 1953, 1954). La superposition des styles se fait sentir jusque dans la région d'Oslo.

La succession et superposition des divers styles permettent une classification rationnelle des zones et avec cela une comparaison plus fructueuse entre les différentes chaînes. Une telle classification n'embrasse pas seulement les données de la pétrographie, mais aussi celles de la tectonique et de la cinématique.

Les surfaces de décollement peuvent se trouver: a) à l'intérieur de la série de couverture; elles peuvent y jouer un rôle important pour la tectonique disharmonique; b) à la base de la couverture (éventuellement au-dessus ou au-dessous d'un étage intermédiaire); c) dans les roches du socle. Les surfaces de décollement sont le plus souvent plissées à plusieurs reprises, ainsi que les surfaces de charriage.

En essayant de placer les zones de la chaîne calédonienne dans ce climat, nous distinguons:

1) zones externes avec série de couverture décollée, socle enroulant tardivement les plans de glissement et la série charriée (exemple: Gjövik, SKJESETH, 1953, 1954).

2) zones plus internes: série de couverture et série intermédiaire décollée et charriée avant le métamorphisme et la mobilisation géo chimique; enroulements multiples; plans de décollement dans le socle probable.

3) zones caractérisées par des plans de décollement du socle; ultramétamorphisme en profondeur, glissements dans les étages supérieurs.

Les matériaux glissés peuvent montrer: a) peu de déformations internes subies par l'ensemble de la masse (exemple du socle: Jotunheimen, de la couverture: Smöla); b) des déformations intimes visibles partout souvent cicatrisées par la recristallisation.

Les événements ne se suivent pas partout de la même façon. *Les zones externes montrent, pour ainsi dire, des stades du début de l'évolution modifiés ultérieurement dans les parties plus internes.* Les zones internes ont passé, au moins en partie, par des stades semblables, mais de nombreuses phases se sont superposées aux formes du début. Les enregistrements compliqués sont la règle, même si les apparences sont parfois simples dues à l'homogénéisation et à la recristallisation.

Une telle évolution est particulièrement bien visible dans les régions à forte pente axiale. Celles-ci présentent de nombreux avantages pour les recherches tectoniques et pétrographiques. L'interdépendance des différents étages tectoniques y devient visible par l'inventaire des données géométriques. Dans les régions surélevées ou déprimées les rapports et interactions des étages ne peuvent être déterminés qu' indirectement. Ils sont le résultat de déductions à partir de certains indices. Le faciès minéral assez homogène et la composition passablement monotone des gneiss de l'Ouest (soulignée par de nombreux auteurs) a longtemps fait penser que le bouclier devait avoir une origine unique. Ce *faciès minéral n'a enregistré que les conditions physico-chimiques régnant lors de l'enfoncement calédonien.* Il correspond à une tendance d'homogénéisation.

Les différences de composition sont dues: soit à la différence de réactivité, soit à des zones de déformation et de transformation plus jeunes, survenues à un niveau plus élevé.

A la monotonie souvent mentionnée dans la littérature s'oppose une multitude de types structuraux et cela dans tous les ordres de grandeur, depuis la texture des roches jusqu'aux grands ensembles. En parcourant les vallées et les fjords, on est étonné de cette richesse dont très peu de données ont été employées jusqu'à maintenant.

On pourrait comparer l'aspect de cette région à une image dont les couleurs sont un peu passées, mais dont le dessin est conservé. Les couleurs correspondraient à la diversité des roches, le dessin aux formes structurales. *Ce dessin a conservé de nombreux enregistrements perdus dans la composition des roches.* Leur nature actuelle est surtout le résultat des migrations déclenchées par la phase d'enfoncement calédonien. Ces migrations observées de 1924 à 1926 (WEGMANN, 1935) ont fait l'objet de belles études par STRAND (1949, 1951).

Beaucoup d'auteurs faisaient venir et font encore venir les granites et les fluides granitisants des profondeurs inconnues. Grâce aux culminations axiales, des espaces autrefois profondément enfoncés sont mis à jour par l'érosion, et nous présentent, à l'état figé, des images montrant les échanges et transformations des roches. J. J. SEDERHOLM avait prévu ces faits quand il a écrit que nous avons ici le privilège de pouvoir jeter un regard dans les «forges de Pluton» (SEDERHOLM, 1907, p. 35).

Le problème des rapakivis va nous fournir un autre exemple de la manière permettant de compléter les renseignements physico-chimiques par des données spatiales et par l'analyse cinématique.

Remarques sur la série des rapakivis.

Il semble utile de mentionner ce problème (sans pouvoir le résoudre pour le moment) en tenant compte des observations faites en Finlande et au Grönland méridional. Les conditions de la descente axiale de la Driva semblent pouvoir ouvrir une nouvelle voie pour les recherches.

On distinguera plusieurs groupes d'hypothèses: celles qui s'occupent de la genèse de ces roches et celles qui traitent leurs rapports avec la chaîne calédonienne scandinave.

En ce qui concerne la *genèse* des granites à structure de rapakivi, deux hypothèses principales s'opposent en ce moment: a) certains auteurs admettent que ces structures ne peuvent prendre naissance que dans un bain silicaté fondu, b) d'autres les considèrent comme dues à la transformation de roches préexistantes, soit de sédiments (BACKLUND), soit de différents matériaux du socle cristallin (WEGMANN, 1938).

Parmi les partisans des bains silicatés fondus, deux courants se font sentir a) un groupe qui rattache la genèse des structures aux conditions spéciales de la solidification du magma (WAHL 1925) et b) un groupe affirmant que la structure dépend de la composition chimique initiale du magma (magma rapakivitique, NIGGLI). Les magmatistes et les transformistes peuvent citer, chacun pour son compte, des observations qui semblent justifier leur manière de voir. Les uns souligneront surtout l'importance des observations faites par J. J. SEDERHOLM, au contact du massif de Viborg, à Buckholm et Bergholm, dans l'archipel de Pellinge (SEDERHOLM, 1923, p. 75 à 92, fig. 28, 29, 39, 40). Les ovoïdes feldspatiques ne s'observent pas seulement dans la masse granitique, mais aussi dans la roche encaissante. Cette observation fut maintes fois contrôlée par des savants venus du monde entier, entre autres par les participants au congrès de 1931.

Des conditions semblables, à une échelle beaucoup plus grande, peuvent être observées au Grönland méridional (WEGMANN, 1938, p. 98 à 121, fig. 59, 60). Dans le district du Cap Farvel par exemple, des traînées noires (graphiteuses et pyriteuses) de la roche encaissante se continuent dans les rapakivis massifs; ailleurs, des ovoïdes nagent dans la roche encaissante, ce qui semble indiquer, au moins pour ces localités, que le granite peut se substituer à la roche préexistante en conservant certains éléments structuraux.

Les magmatistes se plaisent à souligner les faits suivants: les rapakivis passent, à plusieurs endroits, à des porphyres quartzifères comme c'est par exemple le cas entre Kotka et l'île de Hogland. Tous les stades de transition ne sont pas visibles, mais la continuité du phénomène n'a jamais été sérieusement contestée (KRANCK, 1928). Nous avons trouvé dans la région de Pellinge des filons de roches montrant des caractères de porphyres quartzifères et de rapakivis.

Les magmatistes et les transformistes ont souvent sélectionné, chacun pour sa part, le groupe de faits favorables en le présentant comme le seul ayant une certaine importance. Deux sortes de phénomènes ont été constatés par de nombreux observateurs dans les mêmes massifs, à savoir: les contacts intrusifs et les contacts par transformation. On est donc forcé d'admettre les deux groupes de faits. Si on veut sortir de cette contradiction, on doit, en plus des arguments parlant en faveur des hypothèses citées, tenir compte de nombreux autres faits situés en dehors du champ de gravitation limité des discussions entre magmatistes et transformistes. Il serait préférable d'abandonner cette antithèse parce qu'elle est un obstacle à la solution du problème. On arrive à se demander si les questions posées à la nature ont été formulées de telle façon qu'elle puisse y répondre. Ne faudrait-il pas adresser d'autres questions à la nature?

En procédant ainsi, il serait nécessaire de tenir compte de *tous* les caractères des rapakivis, et non pas seulement d'une sélection. Sans sortir du cadre restreint de ce rapport, on peut se poser la question suivante: la structure rapakivi représente-t-elle l'aboutissement d'une évolution vers un équilibre, ou n'est-elle qu'un des états de transition figés sous certaines conditions? Une telle question pourrait être le point de départ d'une série de recherches. Leur orientation serait différente de celle des discussions centrées sur l'existence ou l'inexistence des magmas. Il faudrait pour cela tenir compte des remarquables travaux de BORIS POPOFF (1927) et de nombreux autres faits. Mentionnons deux données comme exemples: a) la structure rapakivi ne se trouve que dans une enveloppe touchant au contact extérieur du massif granitique de Helene Havn, sur l'île de Nunarsuit, au Grönland méridional (WEGMANN, 1938, p. 98—105, fig. 51—53). Elles passent à d'autres variétés granitiques à l'intérieur du massif, entre autres aux roches appelées «prick-granit» par WAHL (1925). b) J. J. SEDERHOLM a observé la coexistence d'olivine et de quartz bipyramidé dans des roches rapakivi du Sud de la Russie. Il est plus que probable qu'il ne s'agit pas d'un équilibre physico-chimique.

On peut affirmer, sans exagérer, que les conditions de formation ne se retrouvent que sporadiquement dans les temps et les espaces géologiques. Il a fallu qu'un concours de circonstances particulières soit réalisé pour donner lieu à cette structure. Une période semble avoir été particulièrement favorable pour un espace allant depuis le sud de la Russie jusqu'en Finlande et en Suède. Pour ces gisements, il existe une certaine unité de temps et de position.

Les gneiss oeillés de la Driva et d'autres gisements analogues ont été considérés comme revêtant le faciès minéral amphibolitique, ce qui ne veut pas dire que les roches de départ avaient également atteint un équilibre. Le faciès amphibolitique (ROSENQVIST) correspond à *la recristallisation* accompagnant et suivant la déformation. Cet équilibre (s'il existe) peut nous donner de précieux renseignements sur les conditions pendant cet épisode, mais il est douteux que ce moment corresponde aussi à la mise en place de ces roches.

En considérant, avec C. W. CARSTENS (1924, 1925), les *gneiss oeillés comme des produits de déformation,* le problème est divisé en deux: 1) la mise en place des produits de départ, 2) leur déformation et transformation. En séparant ainsi leur histoire en deux groupes de phénomènes, on les classe parmi les roches polymétamorphiques.

Ceci nous amène à toucher une autre question: les gneiss oeillés sont-ils d'âge calédonien au non? La notion de calédonien est quelque peu changeante suivant les auteurs. En considérant la série des gneiss oeillés comme des roches polymétamorphiques, on assimilera leurs déformations et leur recristallisation aux mouvements calédoniens.

La mise en place des produits de départ est plus ancienne. Elle se rattache à celle des Helleskiffer, qui eux, sont plus jeunes que le socle. Ce sont naturellement des limites chronologiques assez vagues, mais il est possible de les serrer.

On peut choisir entre deux solutions: a) en admettant que tous les granites rapakivis et les autres roches qui les accompagnent, comme par exemple les anorthosites, se soient formés pendant la même période, depuis le Sud de la Russie jusqu'à la Driva, on doit assimiler les quartzites à une formation plus ancienne que l'Eocambrien. b) en admettant que les Helleskiffer correspondent aux sparagmites, on doit en tirer la conclusion que les structures rapakivis se sont formées à deux reprises en Fennoscandie.

Les massifs de Finlande, de Suède et du sud de la Russie, ont fourni jusqu'à maintenant une multitude de précieux renseignements au point de vue pétrographique, mais leur dissection topographique ne s'étend qu'à quelques centaines de mètres de dénivellation. Ces massifs ne sont donc pas favorables aux recherches poussées vers la profondeur (KANERVA, 1928). En Norvège et au Grönland méridional, la superposition des étages tectoniques a été bousculée de sorte que *la surface topographique actuelle recoupe des niveaux anciens situés* à *l'origine à des profondeurs très différentes.* Les différences peuvent être estimées par l'étude des plongements axiaux. Ce fait livre à nos recherches des *coupures comprenant des milliers de mètres* de l'ancien corps de la montagne. Il est donc possible d'étudier le problème des rapakivis dans une dimension inaccessible dans les massifs classiques de la Fennoscandie. Avant de pouvoir se livrer à ces investigations, il faut pouvoir soustraire les effets des déformations et transformations calédoniennes. De cette façon, il sera possible d'étudier l'origine et le cheminement des agents responsables de la formation des rapakivis. Si nous parlons d'origine, nous entendons par là surtout la détermination des lieux dans l'espace et les chemins suivis, car quelle que soit l'hypothèse, les matériaux, magmatiques ou métamorphisants, doivent partir et s'acheminer depuis des parties fournisseuses de l'écorce vers leur emplacement actuel. Les études tectoniques peuvent ainsi contribuer à la solution d'un problème important et de questions qui s'y rattachent, comme par exemple celles du rassemblement, du cheminement et de la distribution du fluor. Les rapports entre les étages tectoniques et métamorphiques peuvent être reconstruits par des méthodes géométriques et cinématiques, en donnant un support spatial aux raisonnements algébriques de la chimie physique des pétrographes.

Quelques conséquences et interprétations tectoniques.

Les conditions tectoniques de la vallée de la Driva ont été interprétées de différentes manières:

Törnebohm a considéré les répétitions tectoniques comme des écailles charriées vers l'W (Törnebohm, 1896, p. 171). Bjørlykke (1905, p. 404/405) a observé des axes de directions SW—NE et un autre système perpendiculaire plongeant vers le SE. Il pense qu'il y a eu «lutte entre deux systèmes de plissements» au voisinage du vieux massif résistant. Il note que les charriages de Törnebohm sont basés sur l'hypothèse que les gneiss oeillés font partie du socle.

Nous avons maintenu, depuis 1924, une autre interprétation. Les contrôles faits en 1957 semblent confirmer notre manière de voir. Nous admettons une direction axiale moyenne des plis ESE et une plongée sous les terrains de la dépression axiale de Trondheim. Le transport des masses supérieures dans le secteur de la Driva est dirigé vers le SSW. Des directions axiales semblables continuent dans les montagnes de Dovre. Elles oscillent autour de l'horizontale dans le secteur entre Snöhätta et Dovre, comme le montre l'intéressante carte de Du Dresnay (1950). Les axes se relèvent ensuite vers le district de Lesja. Entre Lesja et Vaage, les deux systèmes de déformation continuent de la même façon. Vers le lac de Vaage, un pli du socle enroule l'ensemble des éléments structuraux superposés avec un plongement axial vers l'E. (cf. Björlykke, 1905, p. 406 et 411, Strand 1951) L'encapuchonnement est visiblement tardif et correspond aux autres mouvements de ce genre sur le pourtour et à l'intérieur du bouclier. Törnebohm (1896, p. 172) avait remarqué ces complications, mais ne pouvait pas les expliquer.

L'abaissement axial se fait donc par paliers, un premier, près de Vaage et Lesja, un second, le long de la Driva et du bassin de l'Orkla. La direction axiale oscille entre ESE et ENE. Des flexures, plissements et failles plus ou moins perpendiculaires se retrouvent dans les zones d'inclinaison axiale prononcée.

On pourrait ajouter à ces deux paliers, un troisième, au nord du fjord de Trondheim. Il correspond à un ensellement axial avec une série de complications.

Les axes remontent ensuite vers la culmination des Lierne—Namsos. Ils montent vers le NE dans la zone de Snaasa et vers le NNE dans la zone complexe de Hegsjöfjell (cf. aussi G. Holmsen, 1919). Le complexe charrié métamorphique de la Gula s'élève en l'air, entouré d'une belle auge periclinale. La courbure de cette auge n'est pas un synclinal ordinaire, mais le résultat de l'encapuchonnement des différents éléments tectoniques superposés (cf. p. 51, 53).

En essayant d'obtenir une vue d'ensemble des directions axiales, on s'aperçoit qu'elles dessinent un *feston couvrant la dépression de Trondheim et du Jämtland.* Ce feston tourne depuis les directions NS à EW. En détail, les directions des différents étages tectoniques sont évidemment plus variables. Elles ne datent pas des mêmes phases cinématiques, puisqu'elles comprennent des plissements, des charriages et des encapuchonnements.

En admettant le profil axial et le tracé en festons pour ce secteur de la chaîne, on obtient une autre image de l'articulation des éléments calédoniens. *La limite extérieure de la chaîne, dirigée NE—SW, ne* *correspond pas à son anatomie interne.* Celle-ci est caractérisée par un certain nombre de festons ou guirlandes qui se relaient et se recoupent. Chacun des festons correspond à une dépression axiale majeure. Un dessin schématique a été donné ailleurs dans un cadre plus étendu (Wegmann, 1948). Les régions où deux festons se recoupent et se superposent sont particulièrement intéressantes, mais plus difficiles à étudier à cause de l'interférence des structures. Il est donc préférable de commencer l'étude par des secteurs où la superposition n'est pas multiple et où l'évolution est marquée par un certain nombre de phases principales.

Il ne sera pas possible d'examiner ici toutes les conséquences, tectoniques, stratigraphiques, pétrographiques et géochimiques qu'entraînent ces vues. Nous nous bornerons à citer quelques exemples. On peut les diviser suivant leur ordre de grandeur: 1) en grands blocs ou champs animés de mouvements semblables; 2) en éléments composant la mosaïque dans ces champs; ils correspondent aux déformations qui se greffent sur les mouvements d'ensemble; 3) la structure interne des éléments mentionnés, caractérisant des districts, des affleurements, des échantillons et des préparations.

Les champs ou grands blocs.

Il est possible de subdiviser la chaîne en un certain nombre de champs ou grands blocs qui restent, au moins pendant un certain temps, des unités. Ils sont caractérisés par leur tendance positive ou négative. Dans le secteur qui nous occupe ils se groupent en plusieurs rangées. Les champs sont caractérisés par l'altitude approximative à laquelle se trouve le socle .Même si cette altitude n'est souvent pas facile à déterminer, les différences sont d'un tel ordre de grandeur qu'une confusion n'est guère possible (fig. 4).

La *rangée centrale* est composée par les éléments suivants (du Nord au Sud): le bord septentrional de la dépression de Trondheim et le massif de Namsos ont été pendant longtemps un champ montant. Des restes de matériaux calédoniens et dévoniens sont pincés entre les éléments du socle. La plus grande partie des formations de couverture a disparu par érosion et par glissement. Plus au sud, vient la dépression de Trondheim et du Jämtland. Elle a été remplie par des sédiments précambriens, et des masses glissées cambrosiluriennes et de grandes

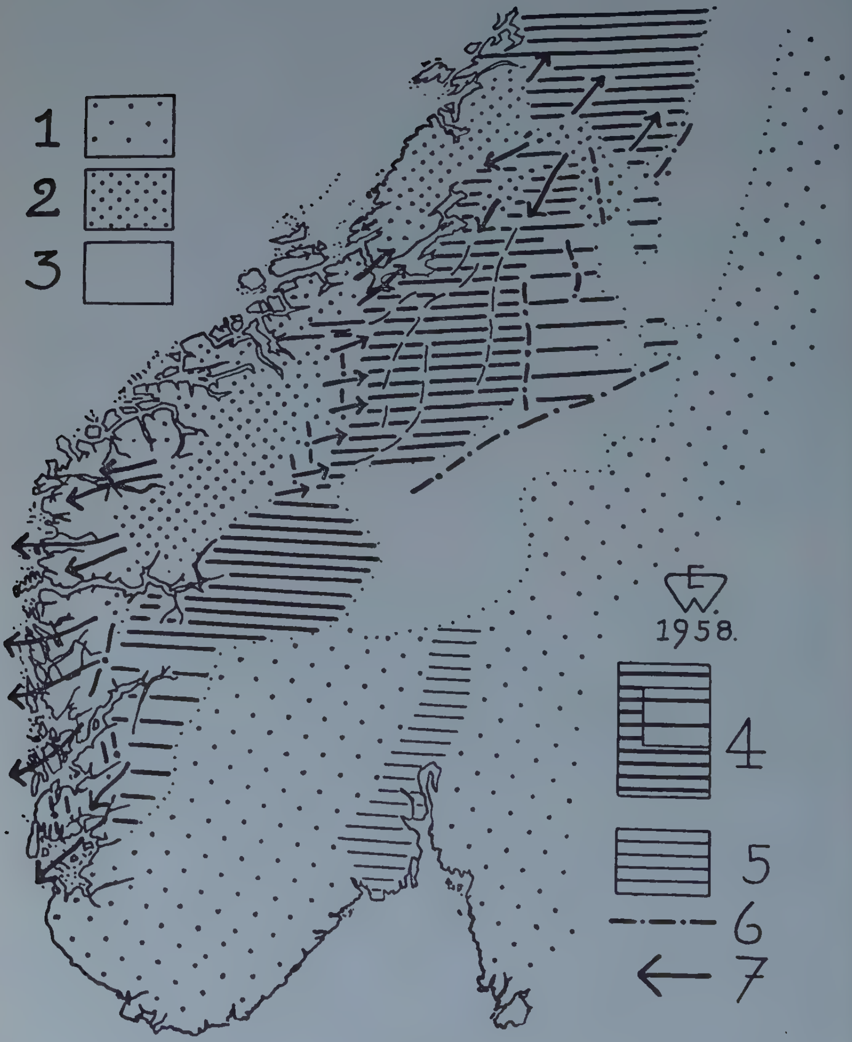


Fig. 4.

quantités de roches volcaniques. Ce remplissage témoigne d'une tendance à l'affaissement pendant la période de dépôt et de déformation. Les matériaux sédimentaires ont été livrés par des régions en voie de soulèvement, comme en témoignent les formations de flysch de divers âges (VOGT ,1928, 1945). Les blocs sont séparés par des zones de failles qui représenteront par la suite des anisotropies importantes et des phénomènes pétrogénétiques de première importance.

Les blocs soulevés n'ont pas seulement livré des matériaux sédimentaires, mais aussi des masses de glissement tectonique d'un tonnage et d'un tirant variables. Des formations paléozoïques de degrés métamorphiques différents ont d'abord glissé les unes sur les autres. Ce premier glissement s'est probablement fait quand ces masses étaient encore situées hors du bassin de Trondheim actuel. L'empilement a rempli par la suite la dépression de Trondheim et du Jämtland en train de s'approfondir. Cet empilement de séries stratigraphiques, volcaniques, et tectoniques a subi le déplacement de la zone déprimée. La dépression a recueilli les sédiments de la fin du cycle calédonien (VOGT, 1928, 1945). L'ensemble fut subdivisé par la suite par des encapuchonnements postérieurs (cf. WEGMANN, 1926). Ces enroulements ont des directions longitudinales, transversales et obliques.

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Fig. 4. *Esquisse schématique de la partie méridionale des Calédonides Scandinaves.*

1) Socle précambrien peu influencé par les déformations calédoniennes.

2) Socle précambrien plissé, écaillé et transformé, avec pincées, parfois étendues, de formations calédoniennes; déformations multiples, décollements et enroulements.

3) Régions où le socle est relativement peu déprimé (ou soulevé), par rapport aux dépressions profondes; couverture décollée, socle bombé et écaillé. Formations sparagmitiques et cambro-siluriennes.

4) dépressions remplies par des masses sédimentaires, volcaniques ou de glissement tectonique, l'ensemble plissé et encapuchonné. Dépression des Jotunheimen, de Trondheim et du Nordland. Dans le cadre: Masses débordantes sur l'Avant-Pays depuis les dépressions intérieures: Vestland méridional, Jämtland, etc).

5) fossé d'Oslo.

6) Flexures axiales de la zone de Vaage—Driva—Orkla, Vestland, en partie disposées en échelon; Nervures méridiennes: Olden, Mulfjell, Sylene; zone des apparitions du socle: Atnesjøen—Hede.

7) Inclinaisons axiales (directions régionales).

Plus au sud, dans la même rangée, s'étend le champ des sparagmites. Le soubassement de la série sédimentaire s'est enfoncé pendant la période de dépôt. Sa couverture décollée et déplacée vers le Sud est actuellement percée à plusieurs endroits par des bombements ou par des écailles du socle encapuchonnant les plans de glissement (Gjövik, Hede, Atnesjöen, Olden etc.). La profondeur du socle est de ce fait assez variable, mais en général plus modeste que dans le bassin de Trondheim.

Une *rangée plus occidentale* comprend: a) les gneiss du bouclier occidental de la Norvège avec leurs pincées de matériaux calédoniens b) au sud, la dépression dans laquelle se sont logées les nappes de Jotunheimen. Ce bassin fut appelé par V. M. GOLDSCHMIDT «fossé de plissement» (Faltungsgraben, GOLDSCHMIDT, 1912). Cette dépression, peu caractérisée pendant la période de sédimentation, s'approfondit par la suite. Elle fit appel à des accumulations de flysch (Valdres-sparagmit) et, les différences de niveaux allant en s'accentuant, aux masses glissées des Jotunheimen. Celles-ci venaient des parties autrefois soulevées, mais actuellement abaissées du champ situé plus au nord. c) plus au sud, s'élève le bouclier du sud de la Norvège; il est coupé dans des formations précambriennes de divers âges et a une histoire tectonique très compliquée.

Une *troisième rangée* de champs est amorcée dans les régions côtières de la Norvège occidentale. Elle est caractérisée par une descente axiale générale vers l'Ouest (cf. les remarquables travaux de REUSCH, C. F. et N. H. KOLDERUP). Cette flexure axiale, beaucoup plus prononcée que celle de la Driva—Orkla, est aussi passablement plus compliquée (paliers et effets accompagnant les flexures). La rangée de champs elle-même a disparu sous les eaux, mais les amorces permettent de se faire une idée sur les grandes lignes.

Les *champs situés au Nord* de la culmination Lierne—Namsos sont disposés d'une façon différente par rapport au grain du socle. Dans les rangées mentionnées avant, les dislocations du socle et des couvertures recoupent perpendiculairement ou obliquement la disposition des rangées. Dans la dépression axiale du Nordland, elles sont grosso-modo parallèles. A partir de la culmination d'Ofoten et de Rombak, les plis tournent de nouveau jusque dans la péninsule de Varanger. Ils recoupent de ce fait les rangées de cette région. Nous ne nous en occuperons pas pour le moment.

Revenons aux champs situés au SW: les aires élevées et déprimées semblent former une *espèce d'échiquier* en ce sens qu'à côté de la dépression de Trondheim s'élève le bouclier de l'W; à côté du champ relativement élevé de la région des sparagmites s'enfonce la dépression remplie par les nappes des Jotunheimen. A côté du bouclier du Sud de la Norvège, se place d'une part, la région d'Oslo et de l'autre, vers la Mer du Nord, un bassin dont seule la marge orientale est visible. Il est possible qu'à côté du champ élevé de Fosen—Namsos et au N du bouclier de l'W, se trouve une aire déprimée qui semble s'annoncer par les affleurements de l'île de Smöla (cf. fig. 4).

Les plissements et les charriages traversent le cadre des champs, ce qui apparaît surtout dans le cas du feston de la dépression de Trondheim.

Cette distribution des tendances de mouvements reflète probablement *les anisotropies et l'évolution des enveloppes profondes de l'écorce terrestre*, inaccessibles aux recherches géologiques directes. Les mouvements de ces enveloppes sont enregistrés indirectement par le soulèvement et l'effondrement des socles et des couvertures. Ce fait permet de les placer dans les échelles chronologiques, mais la réalité matérielle, les roches elles-mêmes qui se déplacent, échappent aux pétrographes et aux géologues. Les recherches géologiques s'adressent à la série des étages tectoniques mue par le coussin profond. Les étages tectoniques semblent se déplacer par rapport aux roches du socle et de la couverture, autrement dit ces roches passent par les différents étages à la cadence des mouvements du champ.

Les déplacements profonds et ceux de l'écorce extérieure sont reliés de plusieurs manières.

L'image des champs se rapproche d'une part de celle donnée par Hans Cloos (1948), d'autre part, elle en diffère par des traits importants. Il semble utile de les mentionner: dans l'image de Cloos, les champs ayant acquis leur forme et leur composition aux temps prégéologiques, évoluent en se déplaçant les uns par rapport aux autres le long des géosutures. Ces dernières jouent un rôle important pour le cheminement des roches plutoniques et volcaniques.

Dans notre perspective, *les champs sont des unités passagères dans l'histoire géologique*. Ils peuvent se souder; la nouvelle unité peut se diviser, et cela d'une autre façon, au cours de l' histoire. La structure en échiquier des temps calédoniens et postcalédoniens semble avoir été groupée en une masse continentale jusqu'à ce que les mouvements tertiaires la recoupent d'une autre façon. Des failles ont taillé dans l'ancienne mosaïque de nouvelles subdivisions très différentes des anciennes. Les mouvements tertiaires créant le bord continental (HOLTEDAHL, 1953) dominent la distribution actuelle des altitudes de la topographie. On rencontre les mêmes problèmes pour les rivages correspondants du Grœnland.

Dans la conception présentée ici, la persistance des limites des champs et des géosutures est moins soulignée; les changements, les regroupements et l'évolution des positions verticales, les déplacements horizontaux et obliques, jouent un rôle important.

Les structures visibles dans les étages tectoniques (accessibles à l'observation directe) apparaissent comme des phénomènes épidermiques, des adaptations passives aux changements en profondeur. Les coussins profonds dont les contours et les mouvements apparaissent comme des fantômes à la surface, ne sont pas accessibles aux investigations pétrographiques et géologiques. Il est évidemment regrettable de constater que le moteur de phénomènes aussi importants se situe hors de l'atteinte des méthodes géologiques. Ces régions profondes sont accessibles aux méthodes géophysiques, mais celles-ci ne peuvent étudier que l'instant actuel. Les états du passé ne sont pas encore accessibles à ces recherches. Ils peuvent être reconstitués par déduction, mais par ce procédé on obtient un grand nombre d'hypothèses diverses parmi lesquelles le choix reste ouvert.

Il est toutefois possible de rassembler un certain nombre d'indices. Ce sont les traces de mouvements réciproques des champs dirigés verticalement, horizontalement ou obliquement. Des traces de ce genre ne manquent pas sur les diverses limites des rangées mentionnées, mais jusqu'à maintenant on ne dispose que d'indices récoltés par-ci par-là. Pour en donner une vue d'ensemble, il serait nécessaire de se concentrer sur l'observation de ces phénomènes avec le but d'en établir l'inventaire.

On est probablement en droit de considérer *les roches basiques et ultrabasiques comme un autre message des profondeurs*. Des intrusions et extrusions basiques sont montées à plusieurs reprises à travers le socle et une partie des couvertures. J. J. SEDERHOLM (1923) a découvert l'avantage que l'on peut tirer de l'observation des filons basiques. Il les a utilisés pour établir une chronologie. D'autre part, en suivant ces repères à travers les coupes profondes des régions à inclinaison axiale, il est possible de relier les structures et déformations des divers étages tectoniques et pétrographiques (Wegmann, 1938, 1953). Les mouvements dans les divers étages peuvent ainsi être distingués par leur style. C'est la meilleure méthode pour synchroniser les mouvements à différentes profondeurs et pour déterminer l'âge des déformations dans des terrains cristallins. Les districts à flexure axiale prononcée de la chaîne calédonienne n'ont pas encore fait l'objet de recherches sous ce point de vue.

Les filons basiques nous font passer aux réalités observables. Les roches des affleurements actuels ont gardé des enregistrements de leur passage par plusieurs étages tectoniques, dans lesquels les différents réseaux de filons basiques fournissent des signaux chronologiques et un quadrillage géométrique. Les déformations du socle peuvent ainsi être reliées à celles des éléments supérieurs. Un champ déprimé accumulera des matériaux provenant des secteurs soulevés. Il appelle des matériaux sédimentaires, volcaniques et tectoniques jusqu'à la limite de sa capacité. A ce moment, la tendance des mouvements verticaux peut se renverser.

Quand l'enfoncement atteint une certaine limite, individuelle et caractéristique pour chaque bassin, le champ commence à remonter et à subir la dégradation et l'érosion. Les matériaux se déversent de nouveau dans d'autres champs, aussi bien en surface par les phénomènes exogènes que sous la surface par les déplacements tectoniques superficiels et profonds. La cadence de ces transferts est variable.

Le domaine des rapports entre les champs et les rangées n'est pas épuisé par ces quelques remarques. Il doit être complété par quelques mentions concernant les phénomènes variés accompagnant le morcellement et la coalescence des blocs.

Les secteurs de montée axiale, comme celui de la Driva, peuvent servir de bons exemples: Lors des transports tectoniques en direction méridienne (à axe plus ou moins EW) le passage de la dépression de Trondheim au bouclier des gneiss de l'W, semble avoir été peu prononcé. La zone s'est affirmée par la suite lors de la formation des flexures, failles et plissements qui s'y rattachent. Les mêmes mouvements et plissements se font sentir plus au sud dans la région du Gudbrandsdalen. Le style de l'ensemble change. Les bords du fossé d'Oslo peuvent appartenir aux mêmes tendances qui progressent à travers le socle vers le S.

Soulignons que le modèle des coussins se déplaçant en profondeur sous l'écorce n'est qu'une des nombreuses images que l'on peut donner d'un phénomène dont il est impossible de toucher les bases matérielles. On peut imaginer de nombreux mécanismes permettant de rendre compte des constatations. Nous avons choisi le modèle mentionné à cause de sa souplesse et parce qu'il ne se prononce pas sur les causes des déplacements.

Les étages tectoniques.

Les *traces observables* dans les échantillons, dans les affleurements et sur les cartes géologiques, appartiennent aux *différents niveaux des étages tectoniques.* Pour les étudier, il faut distinguer deux sortes de mouvements: a) on rassemblera dans un premier groupe les mouvements dans les étages tectoniques et les mouvements reliant plusieurs étages, tous ayant lieu à peu près pendant la même durée de temps. Ce sont *les mouvements plus ou moins synchrones se passant dans différents espaces.* b) le déplacement des étages à travers le tissu rocheux; des traces de différents styles se superposent dans les mêmes matériaux; ce sont les *mouvements métachrones ayant lieu dans le même espace.*

Il n'est pas possible d'exposer ici les méthodes, techniques et résultats de ce vaste domaine. Nous en esquisserons quelques traits en citant des exemples des Calédonides centrales.

A). LE VOLCANISME BASIQUE.

Les filons basiques, les intrusions et les formations volcaniques fournissent, à plusieurs reprises, des repères chronologiques et géométriques. Ils fixent par leur moulage l'image de la fissuration du socle pour une période donnée. Ce réseau permet, dans beaucoup de cas, de reconstituer plus tard les déformations avec leurs composantes verticale et horizontale. Les roches basiques avec leurs laves en forme de coussin (pillow-lava) ont été rencontrées dans le groupe de Röros, le long du chemin de fer, en aval de Kongsvoll. Les couches épaisses de la formation de Bymark, si bien étudiées par C. W. CARSTENS, en sont un autre exemple. Dans les formations volcaniques, on ne rencontre que rarement des filons ayant pu fonctionner comme canaux nourriciers. La plupart des filons est probablement d'importance locale, mais les zones de grandes failles fonctionnant comme charnières (Hinge-Lines) entre les champs, ont probablement ouvert le chemin aux grands canaux nourriciers.

Les glissements (profonds et superficiels) ont coupé les formations volcaniques de leurs racines situées plus au NW que leur position actuelle. L'étude des réseaux de filons basiques dans les parties du socle situées au bord de l'océan pourrait fournir d' utiles informations.

B). LES GLISSEMENTS.

Les glissements des formations de couverture et ceux des parties décollées du socle sur les pentes des géotumeurs sont connus. Ils se font d'une façon différente à la surface et dans les étages profonds. Le style de déformation et l'architecture des empilements en portent le témoignage.

Le lambeau de la Gula (WEGMANN, 1926) s'élève vers le N de sorte que son contact anormal devient bien visible dans le massif du Hegsjöfjell. Les terrains du lambeau se distinguent par un métamorphisme plus accusé et de nombreuses intrusions granitiques. Les couches de la série inférieure, moins métamorphiques, dessinent une belle auge périclinale dans laquelle le lambeau est logé (cf. carte de HOLMSEN, 1919). La série inférieure semble à son tour être décollée et provenir d'un soubassement plus interne. Dans la zone de la Driva les terrains du bassin de Trondheim sont charriés sur les écailles des Helleskiffer (HOLMSEN, 1955).

On distingue donc au moins trois groupes de séries glissées depuis l'intérieur de la chaîne; ces séries sont actuellement superposées. Le lambeau de la Gula a été placé sur son substratum actuel après une première installation des zones métamorphiques. Une seconde adaptation a eu lieu à un moment plus tardif. La superposition et le transport des divers groupes d'éléments tectoniques ont donc eu lieu à des moments différents.

Les remplissages sédimentaires, volcaniques et tectoniques, ont changé les conditions dans les régions chargées et déchargées. Ce changement se manifeste d'une part dans la distribution du métamorphisme et d'autre part dans celle des phénomènes de mobilisation profonde. Parmi ces derniers, on distinguera deux groupes: a) les diffusions à travers les tissus rocheux; mobilisation des composés

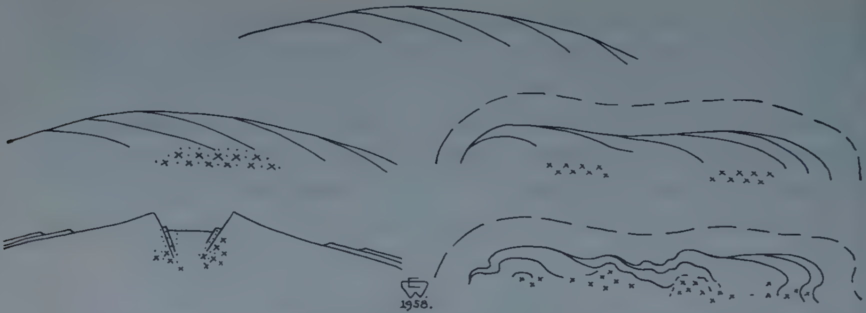

Fig. 5. Deux types d'évolution des plis de fond ou géotumeurs; à gauche: Socle sous une couverture peu épaisse. Mobilisation en profondeur avec séparation partielle des phases fluides; volcanisme explosif; transformations. Fossés avec lèvres soulevées. A droite: Socle sous une masse recouvrante épaisse (composée de sédiments, roches volcaniques et de masses transportées par glissement tectonique). Roches mobilisées et transformées passant par endroits au diapirisme. Les structures antérieures de divers âges sont enroulées.

chimiques; b) les diminutions de cohésion des grains de la roche par l'augmentation des pellicules intergranulaires, ce qui correspond grossièrement parlant, au phénomène inverse de la cimentation d'un grès. Les grains perdant leur cohésion deviennent mobiles les uns par rapport aux autres. Le tout se passe un peu comme dans un sable. Des ensembles de ce genre deviennent macroplastiques. Ils peuvent s'introduire dans les fentes du toit et dans les terrains environnants en formant des diapirs (Wegmann, 1929).

Les terrains calédoniens et ceux du socle montrent de nombreux phénomènes de diffusion (Strand, 1949), de diapirisme et d'intrusion. Presque tous les types de transition et de combinaison sont représentés. Ces différents cas de mobilisation et les structures qui les accompagnent, ont été figés à tous les stades dans les régions gneissiques à l'W et au N du bassin de Trondheim. *Leur fixation prématurée nous livre, pour ainsi dire, un film des événements, si on prend la peine de rassembler ces stades en séries d'évolution.* Pour suivre la genèse des granites et des gneiss, on n'a plus besoin de faire appel aux profondeurs inconnues et à des mécanismes imaginaires. Les «forges de Pluton» ont été mises à jour et sont visibles dans les admirables affleurements polis par les glaciers.

Un autre phénomène doit être relié à la mobilisation en profondeur: si dans un bouclier soulevé les parties inférieures deviennent mobiles,

les conditions de stabilité changeront. Le centre peut s'affaisser, comme c'est souvent le cas. Les bords peuvent être redressés et recouvrir en partie les terrains plus jeunes (cf. fig. 5).

Ces phénomènes de renversement ne sont toutefois pas toujours l'effet d'une mobilisation profonde. Ils peuvent être dûs à la mobilité cataclastique du tissu rocheux, aux mouvements sur les plans de partage et dans les zones broyées. Ceci est le cas dans de nombreux massifs alpins (massifs du Mont-Blanc, de l'Aar, et du St. Gothard). Des conditions semblables règnent dans les massifs cristallins marginaux des Calédonides. Les mouvements de tassement compensant une partie du soulèvement rapide s'effectuent par la protogénisation, la mylonitisation et le glissement sur les joints.

On constate sur les bords des districts gneissiques de Norvège tous les types de transition entre la mobilisation profonde et la mobilité des étages supérieurs.

Les mobilisations profondes et les formes structurales qui les accompagnent peuvent être étudiées dans de nombreuses chaînes. Leur influence ne se fait toutefois souvent sentir que d'une façon indirecte. *Les districts des gneiss de l'W et les régions de Fosen—Namsos laissent apercevoir les rapports géométriques qui sont dans ce cas directement observables.*

Le résultat de ces soulèvements et tassements est dans de nombreux cas un redressement postérieur des structures moins inclinées à l'origine. Ce redressement correspond à des types de structures métachrones.

.C). LES ENCAPUCHONNEMENTS.

Les empilements tectoniques peuvent être plissés ensemble. C'est un premier type d'enroulement. Un des cas les plus fréquents de l'encapuchonnement est le soulèvement tardif du socle sous une masse de charriage. Ces soulèvements subissent souvent des tassements, unilatéraux ou bilatéraux, et forment ainsi des structures en écaille. Ce cas est souvent réalisé dans la chaîne calédonienne: La série des écailles plissées des Helleskiffer est un exemple typique. L'anticlinal de Lönset encapuchonne les séries supérieures. Le lambeau de la Gula et son substratum tectonique sont, les deux ensemble, enroulés par le massif granitique de Tömmeraas (à l'E de Steinkjer). (WEGMANN, 1926).

De nombreux exemples sont visibles dans la péninsule de Fosen, entre le fjord de Trondheim et l'océan. Ils se continuent vers le S jusque dans la région de Molde. Les zones redressées et partiellement renversées, de la vallée inférieure de la Driva, de Surnadal et leur continuation, sont dues à l'enroulement des structures antérieures. Le pli du socle coupé par le lac de Vaage est un exemple typique d'encapuchonnement. Ces plis peuvent passer à des écailles et à des charriages comme le montre la région de Hede, le massif du Mulfjell et le massif de l'Olden. Dans tous ces cas, les mouvements postérieurs du substratum ont enroulés les plans de glissement et de charriage antérieurs, ce qui donne des indications sur la chronologie relative des mouvements.

Les directions des mouvements d'enroulement et celles des charriages antérieurs ne sont pas les mêmes dans les cas généraux. Dans de nombreux cas, elles ont une direction voisine, mais l'angle aigu entre les deux n'apparaît que lors d'une analyse approfondie. Elles donnent lieu à des structures hélicoïdales complexes.

Dans d'autres cas, les structures encapuchonnantes recoupent sous un angle ouvert celles des charriages superposés. Les Calédonides centrales en présentent de nombreux exemples, comme la chaîne des Sylene. Ces recoupements deviennent surtout bien visibles dans les flexures axiales. Aussi bien sur la côte occidentale de la Norvège, que dans la zone Vaage—Driva—Orkla, que dans le pourtour de la culmination des Lierne—Namsos, l'enroulement transversal des plis est bien visible. La flexure Vaage—Driva—Orkla continue probablement en profondeur vers le sud et donne lieu dans les superstructures aux accidents transversaux de la région des Otta (cf. STRAND, 1951, DIETRICHSON, 1950).

L'anisotropie du socle apparaît ainsi à travers le dessin tout différent des formations de surface, et finit par dominer la configuration. Ce dessin est donc fonction de phénomènes ayant leur siège dans divers étages. Ce dessin ne peut être analysé sans la connaissance des régions à flexure axiale où ces étages sont recoupés par la surface actuelle.

Les interférences des étages tectoniques jouent un rôle très important dans les chaînes précambriennes, en Finlande et au Grœnland méridional. Les étages supérieurs y ont disparu sur de vastes régions, ce qui ne veut pas dire qu'ils n'aient pas eu une influence importante sur les structures actuellement conservées. On néglige souvent cette interaction dans les pays où elle n'est pas bien visible.

La répartition des affleurements des étages tectoniques de la chaîne calédonienne est idéale pour faire voir l'interdépendance des étages tectoniques. Elle servira à élucider de nombreuses structures des chaînes précambriennes et à clarifier les mouvements superposés des chaînes alpines.

Les mouvements verticaux peuvent apparaître, soit sous forme de flexures, soit sous forme de failles. Les deux peuvent être combinées. Leur apparition dépend de plusieurs facteurs qui interfèrent. Les plus importants sont: la nature et la structure de la roche, la vitesse de déformation et le sens du déplacement.

Des dislocations lentes permettent aux tissus rocheux de s'adapter. Des changements rapides cisaillent l'écorce du bas en haut. Des exemples impressionnants sont visibles dans beaucoup de fjords du Grœnland méridional. Ils sont un peu voilés par les déformations subséquentes dans les régions du socle calédonien. Les flexures se montrent surtout, d'une part dans les terrains profonds, et d'autre part dans les couvertures. Le mécanisme de la mobilité interne est très différent dans les deux cas.

Les nombreuses observations dans divers étages tectoniques faites dans des chaînes anciennes et récentes, montrent des configurations différentes suivant que la composante horizontale ou verticale est prépondérante. Nous avons relevé de nombreux indices de mouvements de ce genre, mais il n'était pas possible d'en établir un inventaire. Les limites des rangées semblent être particulièrement dotées de ces indices.

Jetons encore un regard sur la distribution et les directions des encapuchonnements calédoniens: on peut distinguer des enroulements longitudinaux et transversaux par rapport aux déformations des éléments tectoniques superposés. D'autre part, puisque ces dernières directions tournent en formant des festons, il est possible de distinguer un réseau avec une direction grosso-modo méridienne et une autre qui la recoupe variant entre E—W et NE—SW.

La culmination des Lierne, le massif de l'Olden, le Mulfjell et les Sylene appartiennent à la série méridienne. Ces soulèvements tardifs sont arrangés en échelons. Puisque les structures des séries charriées décrivent une courbure, elles sont subparallèles au N et presque transversales au S. Les zones Vaage—Driva—Orkla et la zone du Vestland forment deux autres structures méridiennes, en partie aussi arrangées en échelons. D'autres accidents de ce genre se retrouvent plus au nord, traversant la chaîne entre la culmination d'Ofoten (fenêtre de Rombak

—Sjangeli) et la péninsule de Varanger. Entre ces deux secteurs, dans le Nordland et la Laponie suédoise, les dislocations méridiennes sont subparallèles aux éléments charriés et plissés. On peut compter le fossé d'Oslo parmi les éléments subméridiens.

D'autres séries recoupent, à divers angles, les précédentes, telles que les percées du socle depuis Atnesjöen jusque dans la région de Hede. Le style de ces montées et écailles est très différent des structures correspondantes dans la zone de la Driva. Ces derniers encapuchonnements ont eu lieu sous une grande et épaisse masse recouvrante. La comparaison des deux est instructive.

Dans tous ces secteurs la tectonique à superposition multiple est caractéristique. Elle semble avoir été entrevue par REUSCH pour le Vestland.

Les mêmes anisotropies du socle jouent un rôle différent sous la surface et en profondeur où elles sont soumises à l'anatexie différentielle. Les rapports des anisotropies du soubassement avec les fronts de l'ultramétamorphisme permettent d'expliquer de nombreux mouvements rattachés jusqu'ici à des forces et «poussées» hypothétiques. Le rôle changeant des anisotropies pendant les phases montantes et descendantes du socle montre bien l'évolution irréversible et distinctive de chaque compartiment de l'écorce. La suite de ces épisodes donne la base à une classification rationnelle, géométrique et cinématique, des différentes parties de la chaîne.

Les comparaisons des zones avec les Alpes par HOLTEDAHL (1936, 1948) ont été d'une grande utilité. Entre temps, les idées sur la formation des Alpes ont évolué. Elles sont en pleine transformation. De nombreuses tendances, en partie tout à fait opposées, se font jour et essaient de transformer l'image traditionnelle. En basant une nouvelle comparaison sur des caractères géométriques et cinématiques, plutôt que sur des notions compréhensives et dynamiques, on éclairera d'autres facettes des problèmes. Une telle comparaison dépasserait le cadre de ce rapport, parce qu'il faudrait exposer les différentes conceptions alpines et leurs raisons.

Epilogue.

Essayons d'esquisser les traits les plus importants de l'image obtenue par trois années de pérégrinations (1924—26), par l'étude de nombreux travaux et cartes, et par le séjour en 1957: Sous les trains de plis et de charriages apparaît un autre réseau de structures actives. Les festons et guirlandes calédoniens les recouvrent et s'y adaptent de plusieurs façons. On pourrait les comparer aux plis d'une étoffe jetée sur une figure humaine; elle fait deviner le contour du corps et des membres, tantôt en les cachant, tantôt en les mettant en valeur. L'état actuel de la chaîne est comparable à une statue, dont le dessin suggère le mouvement dans plusieurs parties. En déchiffrant les enregistrements, on obtient des données permettant de ramener à la vie les différents membres de la chaîne.

Une telle image donne une idée des conditions géométriques et cinématiques. Si on veut saisir d'autres facettes de la réalité, on doit ajouter les témoins des passages du tissu rocheux par l'échelle des étages tectoniques. Ces passages et la succession des enregistrements leur confèrent une nouvelle tonalité.

Les phénomènes de la surface (érosions, transports, sédimentation) sont déterminés, parfois par les anisotropies profondes, parfois par celles des étages supérieurs, parfois par l'interférence des deux groupes de facteurs.

L'évolution des vastes surfaces continentales est surtout déterminée par le premier groupe. L'évolution des zones orogéniques est avant tout caractérisée par le second groupe. Il est évident que tous les types de la nature sont le résultat de l'interférence des deux groupes, mais dans la majorité des cas, l'un d'entre eux est prépondérant.

Dans les secteurs des Calédonides, la distribution des sparagmites semble être déterminée par des bassins de direction méridienne comme c'est le cas en Norvège méridionale, au Jämtland et au Västerbotten. Les bassins du Finnmark, bien que déformés plus tard dans une autre direction, ont probablement eu, à l'origine, une direction méridienne. Les bassins de sédimentation des formations de flysch, et des conglomérats et grès dévoniens semblent être liés aux éléments de mouvements calédoniens. La répartition des formations volcaniques obéit à l'interférence des deux.

L'analyse des facteurs reliés aux étages tectoniques et aux zones profondes et de leurs rapports fournit une image nuancée et variée. Elle se rapproche plus de la variabilité infinie des phénomènes de la nature que les cadres plus rigides du passé. Le perfectionnement doit probablement être cherché dans le développement de modèles encore plus différenciés. Il sera possible de les organiser de telle façon qu'ils pourront être employés par plusieurs catégories de chercheurs. Les âmes simples peuvent se contenter de ne faire jouer qu'une ou deux des multiples possibilités, tandis que ceux qui désirent se rapprocher de la réalité complexe, essaieront de rassembler les résultats des interférences de plus en plus nombreuses. En tenant compte de ces buts, il est possible de développer les méthodes et techniques actuelles dans la direction voulue. Beaucoup d'entre elles portent l'empreinte des conceptions simplistes qui leur ont donné naissance, et du cadre dans lequel elles devaient servir d'outils. Une conception plus différenciée demandera aussi un équipement et un outillage mentaux plus évolués.

Le tissu rocheux a enregistré les événements de plusieurs étages tectoniques. Les terrains du socle ont passé par deux groupes de trempes, une partie des terrains de couverture par un groupe seulement. La structure dépend des formes et de la grandeur des unités de mouvement qui varient depuis la grandeur du minéral et de la pellicule intergranulaire jusqu'aux champs tectoniques et à leurs zones de suture.

Les formes actuelles sont dues à la superposition successive des enregistrements dans divers étages. Dans la règle, elles ne peuvent pas être expliquées par des mécanismes simples. Les stades successifs du métamorphisme semblent montrer la même tendance.

A côté des structures ayant pris naissance dans un seul étage tectonique, nous avons essayé de mettre en évidence les interférences entre les mouvements des étages voisins. Ces phénomènes peuvent être décelés par des méthodes que nous avons essayé de mettre au point. Pour faire cela, il fallait chercher de bons exemples en passant d'une chaîne à l'autre. C'est ainsi que notre chemin nous a mené en Norvège dans les années 1924 à 1926 et de nouveau en 1957.

C'est avec un sentiment de vive reconnaissance que nous soumettons ces lignes à l'appréciation de nos amis et collègues norvégiens. Nous espérons qu'en éclairant les faits sous un nouvel angle, et d'une façon peut-être un peu personnelle, il sera possible d'apercevoir de nouvelles perspectives. Cette différence d'angle de vue peut être utile même à ceux qui n'adoptent pas notre manière de voir, en mettant en relief un certain nombre de faits qui leur permettront de continuer le chemin qu'ils ont choisi; ce chemin leur offrira une autre perspective, de sorte que le but leur apparaîtra sous un autre angle.

Nous n'avons pas voulu supprimer des problèmes qui semblent être considérés de certains côtés comme résolus. Nous avons préféré en faire le point et les situer dans un vaste cadre. Cette attitude décevra peut-être ceux qui désirent des solutions simples et rapprochées. La conception des superpositions multiples nous fait paraître le but plus éloigné, mais elle indique aussi des chemins d'accès qui ne sont pas des ponts dans les nuages. Souhaitons que ce rapport puisse contribuer à la connaissance approfondie d'une des chaînes les plus fascinantes.

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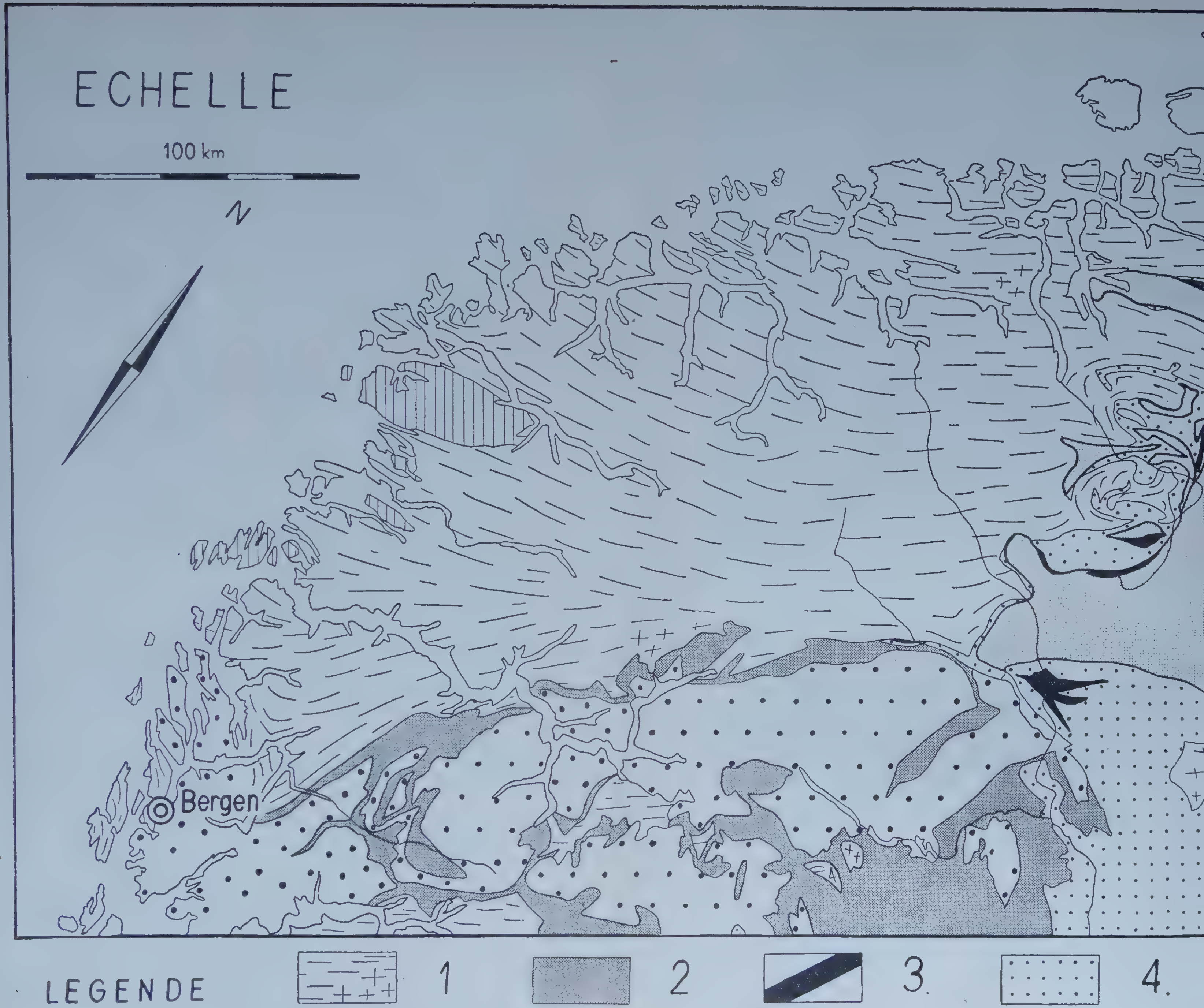
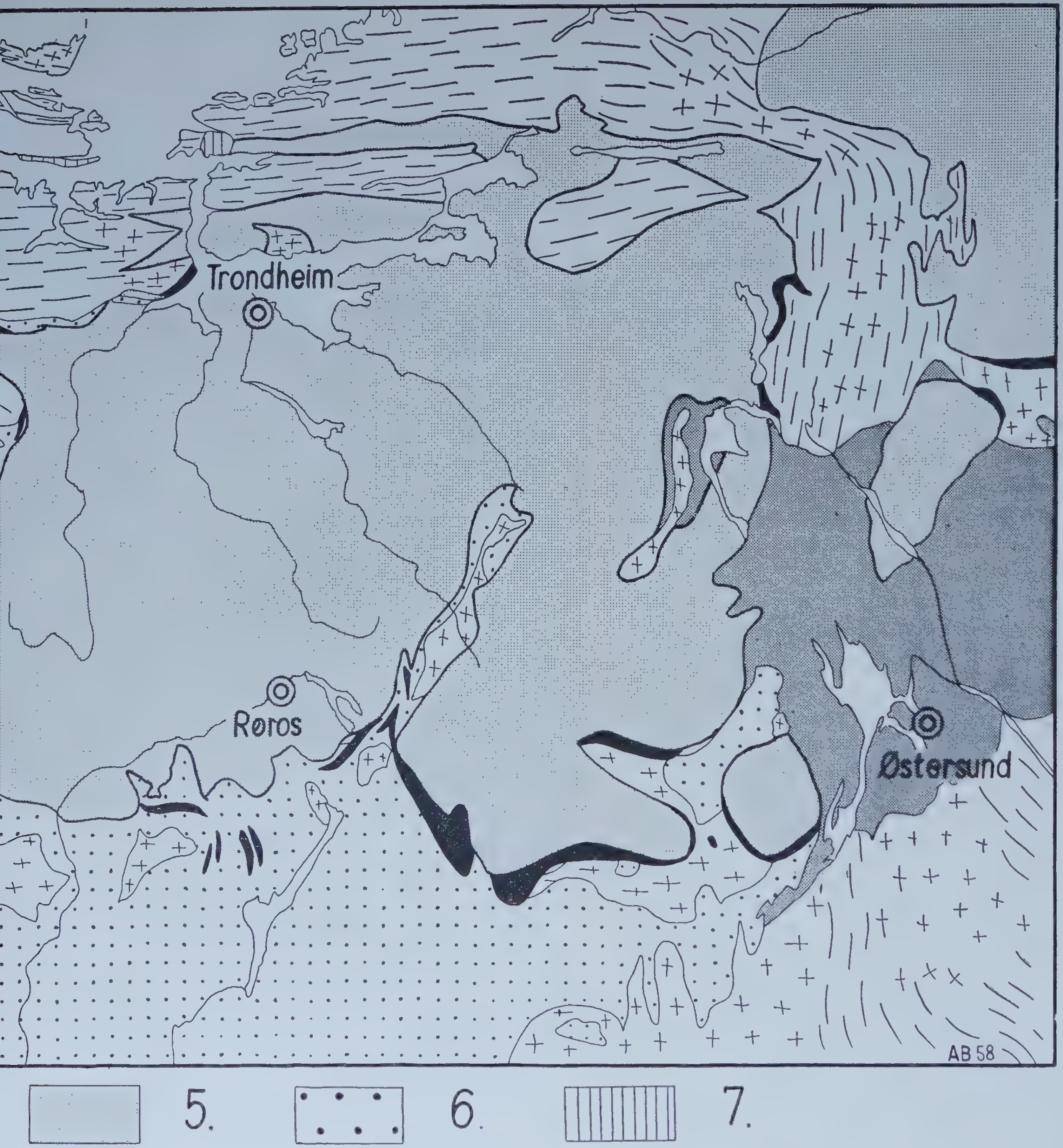


Pl. 1. *Carte tectonique de la chaîne calédonienne entre Bergen et Östersund (schématisée).*
1) Complexe des vieux gneiss avec zones de granite (Jämtland).
2) Cambro-Silurien autochtone.
3) Zone de déformation intense à la base du complexe de Trondh
gneiss oeillés et roches analogues).
4) Couverture du noyau des vieux gneiss (sparagmites).
5) Complexe de Trondheim (y compris intrusions d'âge calédonien



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6) Masses charriées de Jotunheimen.
7) Dévonien (sédiments gréseux et conglomératiques).

Les trainées Cambro-Siluriennes pincées, plissées et charriées de la côte occidentale, ainsi que les formations de la région de Molde, Smøla et Hittra ne sont pas différenciées.

ON THE OCCURRENCE OF TELLURIUM IN NORWEGIAN GALENAS

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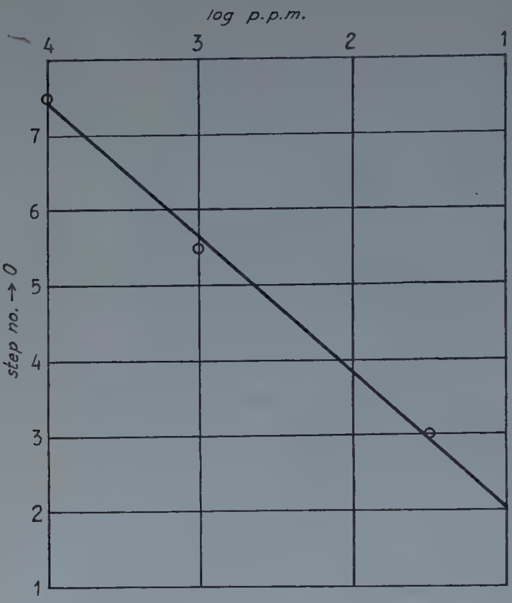
Abstract. Te has been observed spectrographically in but a few of the more common sulphide minerals. Of these galena is probably the most important. Not seldom Te concentrations considerably higher than 10 p.p.m. are found in galena specimens. The highest Te concentrations — more than 100 p.p.m. — are here nearly always accompanied by unusually high contents of Bi and Tl. On the whole the element association Te-Bi-Pb-Tl is very characteristic in sulphidic environment.

Tellurium is a rare element. The average content of Te in the lithosphere is not yet known with any degree of certainty. Based on data given by I. and W. NODDACK (1) (2) V. M. GOLDSCHMIDT (3) quoted the figure 0.0018 p.p.m., but actually he did not believe in this figure. On the contrary, for theoretical reasons he assumed the average Te concentration to be considerably higher. In the iron phases of meteorites he even thought that it might be as high as 10 p.p.m. Nevertheless the Te contents of most common minerals appear to be too low to be detected by direct spectrochemical analysis, i.e. well below 10 p.p.m.

It is well known that Te is highly enriched in certain types of sulphidic ore deposits, in particular in association with Au. The gold (and silver) tellurides are, however, rare minerals, and so are the known tellurides of Bi, Cu, Ni, Pt. Te is also enriched to some extent in pyrrhotite-pentlandite ores and in certain copper ores. From the latter the production of Te metal is obtained. It is also known that rather high Te contents may be found in galena (3).



The present paper reports on the spectrographic examination of a considerable number of Norwegian galenas, and some other minerals. The spectrograms were taken with a Hilger "Large" quartz spectrograph by Mr. J. HYSINGJORD and Mr. F. WOLFF. The conditions of arcing etc. were essentially as described elsewhere (4). The exposure time was kept constant at 90 seconds, and a few not quite successful (weak) spectrograms were left out of consideration. The standard mixtures were made from sylvanite (Offenbanya) assumed to contain about 60% Te, and a Te-poor galena (Bjørnøya, see below) for base substance. In order to obtain fairly accurate estimates of the intensities of the Te line — 2385 — a rotating stepped sector was applied. Thereby the lines were divided into 7 intensity steps, the exposure time for the nth step being $90/2^{n-1}$ seconds. The intensity readings were made by recording the ordinal number of the vanishing step in each case. In this way the working curve shown in the figure was obtained from the standard mixture spectrograms. It is seen that the curve is a nearly straight line, so that one may assume that it will give fairly reliable results.

The results are summarized in the table. Within each of the groups I II III IV the specimens have been arranged according to increasing contents of Te. In a number of cases the contents of Tl, Bi and Sb have been added, these figures having been taken from an earlier paper (5). They are less accurate than the figures for Te, but they are sufficiently accurate to bring out the characteristic features. The determinations of Tl, Bi, Sb and Te have been carried out on the same specimen in all cases. The Ag contents, which have also been deter-

Tellurium contents of Norwegian galenas.

(The figures are logarithms of p.p.m.)

	Te	Tl	Bi	Sb
I. Contact deposits in the Oslo Region.				
Konnerud district (2 localities)	<1			
Grua district	<1			
—»—	<1	<1	<1	2
—»—	<1	<1	3	2
Konnerud district (2 localities)	1.0			
Korsegård, Eiker	1.0	<1	2	2
Grua district	1.0			
—»—	1.3			
Konnerud district (2 localities)	1.7			
—»— (Wedelseie)	2.0	3	4	<1
Grua district (Nyseter)	2.0	<1	3	2
Kjenner Bi Mine (111-cleavage)	2.3			
—»— —»—	3.0	2	4	<1
Ia. Pegmatite, Langesundsfjord (Eikahl.)	2.0	<1	1	2
II. Mineral veins in Southern gneiss areas.				
Tråk, Bamble (2 samples)	<1			
—»—	<1	<1	3	3
Fiskum—Eiker district	<1			
—»—	<1	<1	3	3
—»— (2 localities)	<1	<1	1	3
Lassedalen, Meheia	<1	<1	2	2
Bergggård, Tyristrand	<1	<1	<1	2
Skuterud, Modum	<1	<1	<1	3
Jaren, Hadeland	<1			
Fiskum—Eiker district	1.0	<1	<1	2
—»—	1.0	<1	3	3
Espeland, Vegårdshei	1.3	2	2	4
—»—	1.7	2	4	3
Fiskum—Eiker district	1.7	<1	3	3
Sandåen, Gjerpen (111-cleavage)	2.3	2	4	<1
III. Western Norway.				
Krækkja, Hardangervidda	1.3	<1	3	3
Skjoldevik, Skjold (111-cleavage)	2.3	2	4	2
IV. Northern Norway.				
Svenningdal Ag Mines	<1	1	3	4
Eiterjorden, Beiarn	<1	<1	2	3
Funta, Tysfjord	<1	<1	3	3
Mosbergvik, Balsfjord	<1	2	<1	2
Jakobselv	1.0	<1	<1	2
Bjørkåsen Mines (111-cleavage) (3 sp.)	1.7			
—»— —»— (2 sp.)	2.0			
—»— —»—	2.3	3	4	1
V. Bjørnøya (Russehavna)	<1	<1	<1	3

mined, I have not quoted as they do not show any obvious correlation to the Te contents.

The occurrence of Te in galena does not seem to be markedly dependent on the geological environment. Both in the contact deposits of the Oslo Region and in the various hydrothermal deposits there are Te-poor galenas as well as galenas containing Te in concentrations of up to 200 p.p.m. or more. However, on the whole the galenas of the Oslo Region appear to be on the average somewhat richer in Te than those of the other districts. Relatively few of the former are lower in Te than 10 p.p.m., while in hydrothermal veins etc. such low Te contents are highly predominant. It is seen that the highest Te concentrations, 100 p.p.m. and more, are nearly always found in particularly Bi-rich galenas with octahedral cleavage, and that these galenas invariably exhibit Te concentrations of this order of magnitude. Such galenas are of course relatively very rare. But they occur in contact metasomatic as well as in hydrothermal deposits. Some of the Te-rich galenas do not show octahedral cleavage, e.g. specimens from Konnerud and Espeland, but are still rich in Bi (1000 p.p.m. and more) if not sufficiently rich for the octahedral cleavage to appear (6). The table contains one instance of galena which is rich in Te and poor in (actually nearly free of) Bi, but that comes from nepheline syenite pegmatite, which is, as is well known, very peculiar as to element associations.

Also Tl shows a strong tendency to be enriched along with Te. But this is probably due to the association Tl-Bi which is a very striking feature in particularly Bi-rich galenas.

The table contains one instance of a rather Tl-rich galena which is extremely poor in Te and Bi (Mosbergvik), and one less pronounced case (Espeland 1). This shows that Tl may appear independenty in galena.

Between Te and Sb there is a distinct negative correlation. In particular the very Te- (and Bi-) richest galenas are extremely poor in Sb. In this case too the real negative correlation is probably between Sb and Bi.

Conclusion.

Even if galena appears to be on the average the Te-richest of the commoner minerals its Te content is in the majority of cases lower than 10 p.p.m. In the relatively very rare galenas exhibiting Te contents of 100 p.p.m. and upwards the association with high amounts of Bi and Tl is very characteristic. I am not able to explain this adequately. Crystal chemical principles of course will allow Te as well as Tl and Bi to replace S and Pb respectively in the galena structure, but they will not explain why these elements enter galena in company (except perhaps in the case of Tl and Bi which would to some extent compensate each other electrostatically). Exceptional cases (Langesundsfjord) also show that Te may enter galena independently. The characteristic association Pb-Tl-Bi-Te must be due to geochemical conditions regulating the supply of elements during the formation of these rather unusual Pb ore deposits. The quantities of these elements contained in a parent granite magma may have been largely concentrated in these deposits due to special chemical physical and geological conditions. Since the galenas of the Oslo contact deposits appear to be relatively rich in Te, and since the absolutely Te-richest galenas appear to belong to the very Bi- and Ag-rich type found in certain high temperature deposits (6), there is some reason to believe that the enrichment of Te in galena is promoted by high temperatures. Therefore Te-rich galenas are probably in general of high hydrothermal or pneumatolytic origin.

Other minerals — essentially sulphides — are being examined in this institute for their contents of Te. It may be added here that the association Te-Bi(-Pb) is strikingly apparent also in all examined bismuthinites and "galeno-bismuthinites" (6 Norwegian localities), which exhibit Te concentrations from about 100 p.p.m. to about 5000 p.p.m. (bismuthinite from Bastnäs in Sweden has been shown more than 50 years ago to contain 9500 p.p.m. Te (7)). Also the existence of minerals like tetradymite demonstrates this association. Otherwise there seems to be but few mineral species which may contain Te in concentrations higher than 10 p.p.m. Among these is bornite (Mosnap, Telemark, 30 p.p.m.).

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NOTISER

Native Bismuth in the Molybdenite Deposit at Skjoldevik, Haugesund Peninsula, Western Norway.

BY

IVAR OFTEDAL

During a spectrochemical survey in search for host minerals for tellurium it was found that one molybdenite crystal — from Skjoldevik (1) — contained Te in excess of 100 p.p.m. This rather surprising fact naturally lead to the examination of further molybdenites. Molybdenite samples from 20 deposits in various parts of Norway were searched for Te. However, in none of these Te could be detected spectrochemically. The Skjoldevik sample was then searched for other elements. This at once revealed the presence of strong Bi lines, which were, like the Te lines, absent in all the other molybdenite spectrograms. A spectrogram of a second molybdenite crystal from the same Skjoldevik specimen also exhibited Bi and Te lines, but these were considerably weaker than those recorded in the first case. These observations strongly indicate that the Skjoldevik molybdenite does not contain Te by itself, but is contaminated by varying amounts of some Bi-rich mineral which contains considerable quantities of Te. This mineral could not be observed directly by means of a loupe, so it must be rather finely dispersed in the molybdenite. It is highly probable that this association of molybdenite with Bi and Te is more or less incidental, not due to the presence of exsolution products of a molybdenite originally rich in Bi and Te. This is further confirmed by the following observations.

On careful inspection of the same Skjoldevik specimen it was found to contain, in addition to the abundant molybdenite, small very soft and sectile mineral grains with a bright metallic luster, embedded in a fine grained mass of quartz. The size of these metallic grains or clusters is usually much less than a millimeter. One single grain was much larger, exhibiting a surface of a few square millimeters. The luster of this larger grain is nearly white with a reddish or yellowish tint, while the smaller grains seem to be more greyish. Therefore it was suspected that two different mineral species were represented. However, spectrograms did not reveal any important difference. The only abundant element observed is Bi in both cases. In addition there is some Te, somewhat more in the small grains than in the larger one.

Other components (Ag, Pb, Sb) vary in quantity but are quite subordinate in both cases; these are probably mainly constituents of associated other minerals, e.g. galena. In the same way Mo lines, which are more or less intense in all spectrograms in question, may be explained: they are almost certainly due to contamination by molybdenite. Thus the larger grain must obviously consist of native bismuth. Considering the higher Te content and the greyish colour of the small grains, these might possibly be tetradymite. But X-ray powder photographs have established beyond doubt that the main substance of both large and small grains is one mineral species — native bismuth. The colour difference can be explained by the fact that the small grains are rather rich in Mo, so that their colour may be partly due to molybdenite.

Because of the small quantities available — far smaller than our usual samples for arcing — it was not possible to interpret the spectrograms in terms of quantitative data. But the Te contents of the bismuth are probably of the order of a few per cent, apparently about 5 times higher in the small grains than in the larger one. It is fairly obvious that it is this Te-rich native bismuth which is also found hidden in the molybdenite crystals a few centimeters away.

According to the description by J. Schetelig (1) the molybdenite is found in quartz-rich veins in the vicinity of a granite body. In a vein at the immediate granite contact the Bi-rich galena described by him was found. It is possible that Pb has been introduced locally into a medium containing Bi and Te, thus giving rise to the Bi-rich — and Te-rich(2) — galena. Microscopic examination of this galena has indicated that it has been subjected to exsolution, some of the supposedly exsolved mineral grains being almost certainly native bismuth (3). Since the independent occurrence of native bismuth in the deposit has now been established, it is a possibility that the native bismuth found within galena crystals may be remnants of pre-existing bismuth crystals which have been partly consumed during the formation of the Bi-rich galena.

It is a pleasure to record my gratitude to Dr. H. Neumann and other workers in the Min.-Geol. Museum in Oslo for supplying nearly all of the specimens examined and for checking the identity of the native bismuth by its X-ray powder pattern, and to Mr. Jens Hysingjord for careful preparation of the samples and optical spectrograms.

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Traces of an Eocambrian orogeny in Southern Norway.

BY

NILS SPJELDNÆS

During a trip to the Mjøsa District, Southern Norway, together with Prof. F. Geukens (Louvain, Belgium), studies were made on the boulder content of the Moelv Conglomerate (tillite) in the railway sections south of Moelv.

Prof. Geukens suggested that some of the boulders of Biri Limestone, which are locally abundant in the conglomerate, were metamorphosed before the formation of the Moelv Conglomerate. Later the present author made some detailed studies which confirm Prof. Geukens' suggestion.

The geology of the district, the stratigraphy of the Sparagmite series and a description of the locality are given in HOLTEDAHL (1953), where references to earlier papers are given.

The boulders and stones of Biri Limestone are locally very common in the Moelv tillite. Generally limestone boulders in conglomerates have well preserved structures due to the protection by the matrix, but in this case the limestone boulders show exactly the same signs of pressure as does the limestone exposed near by. Thin bands (1—12 mm) of coarse, recrystallized calcite with "pressure twinning" are common (see fig. 1). The bands generally follow the schistosity of the limestone. In the boulders the recrystallized bands showed no orientation; contrary to what would be expected if they had been formed by deformation in the conglomerate. The distribution of band directions was studied in about 40 boulders. It was random, and showed no relation to the direction of pressure in the conglomerate. Studies on polished surfaces and in thin sections also showed that the recrystallized bands are cut discordantly by the outline of the boulders, thus indicating that they were formed before the sedimentation of the conglomerate. The whole conglomerate has been compressed by later (Caledonian) movements, and the compression cracks found in many boulders are thin, without any signs of recrystallisation, and they cut the bands at varying angles (fig. 2).

This probably indicates that the Biri Limestone was folded before the deposition of the Moelv Conglomerate, and thus that the deposition of the Sparagmites in Southern Norway was interrupted by an orogeny, probably between the Biri Limestone and the Moelv Sparagmite. This is also indicated by the fact that the Sparagmite Series in this region consists of two sedimentary cycles. The older one starts with the Brøttum Sparagmite, a dark graywacke with conglomerate and shale beds. Then follows the Biri Conglomerate, a fluviatile conglomerate with typical "lag bedding" and imbricate boulders. The Biri Limestone,

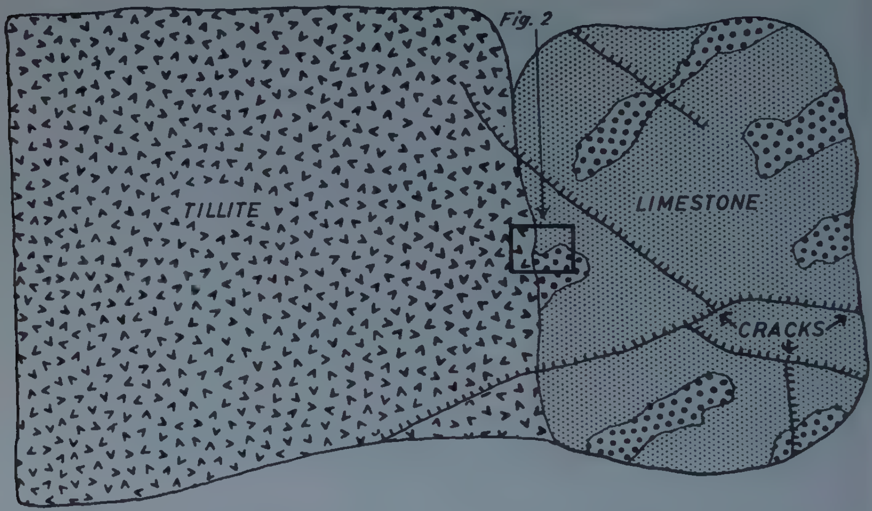

Fig. 1. Drawing of a polished section of a limestone boulder in tillite, showing recrystallized bands (larger dots), and pressure cracks (indented lines), belonging to two different phases of deformation of the limestone. The small square indicates the spot from which fig. 2. is taken. About natural size.

which completes this cycle, consists mainly of gray and black shales, with locally abundant and thick limestones, which have given name to the formation. The younger cycle starts with the Moelv Sparagmite, a typical red arkose consisting of coarse, angular granitic debris. It is followed by the Moelv Conglomerate, a tillite forming in many localities the base of the younger cycle, which has a much wider geographic distribution than the older one. The Moelv Conglomerate is followed by the Ekre Shales, green and red, partly varved shales. They are followed by the Vardal Sparagmite, which is a sandstone with numerous kaolinized feldspars. It grades into the "Quartz Sandstone" (Kvartssandsteinen), which is a typical orthoquartzite. According to the author's observations, no considerable lithological break is found between this formation and the fossiliferous Lower Cambrian.

In Finnmark, Föyn (1937) reports an angular disconformity between the Porsanger Dolomite and the tillites. The angle is low, and the disconformity might be due to tilting rather than to folding of the beds. The boulders of Porsanger Dolomite in the tillites are apparently not metamorphic, but the fact that they consist of dolomite indicates that the dolomitisation took place before the formation of the tillites.

In Normandy tillites are found also (Wegmann, Dangeard & Graindor, 1950) as the "poudingue de Granville". This conglomerate

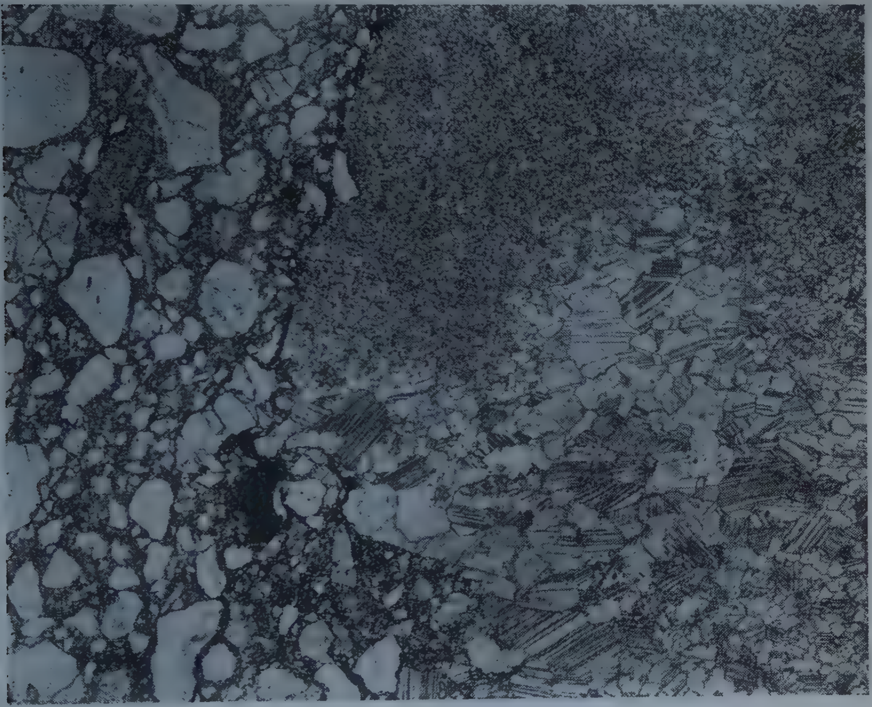

Fig. 2. Photomicrograph of a thin section through the border between tillite and limestone (cf. fig. 1), showing the discordant cutting of the recrystalized band (lower part of fig.), and the "pressure twinning" in the calcite. 50x.

has been regarded as a part of the Brioverien, that is as the basal conglomerate of the upper part of the Brioverien, or as the basal conglomerate of the Cambrian. According to the studies of MATHIEU (1943) the boulders in the conglomerate include metamorphic sediments and a granite which is younger than the lower part of the Brioverien. This indicates that the older rocks were folded and intruded by a granite before the formation of the tillite. This orogeny has been called the "Phase Candomienne" in this region, and has also been referred to the doubtful "Assyntic Orogeny".

Thick lavas (rhyolites) are found between the tillite and the fossiliferous Cambrian.

The observation referred to indicates that the widespread tillites at the base of the Cambrian are also in many areas accompanied by evidence of an orogeny. It has been suggested, that the tillites will be a convenient lower border for the Cambrian, and the presence of an orogeny adds to the value of this suggestion.

The glacial period indicated by the tillites might have been of long duration, and the orogeny might not be strictly contemporaneous in all regions, but, in the absence of fossils, this will be the best available stratigraphical border.

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En magnetitforekomst på Munkholmen ved Trondheim.

An occurrence of magnetite at Munkholmen near Trondheim.

AV

FREDRIK CHR. WOLFF

Summary: A layer rich in magnetite has been found at Munkholmen near Trondheim. The magnetite layer occurs together with blue magnetite bearing quartzite («blåkvarts») in a greenstone, with pillow structure, of the Støren—Bymark-Group. These blue quartzites are quite common in the greenstones of the Støren—Bymark-Group and the Munkholmen occurrence is one of these. Magnetite layers associated with the blue quartzites are not that common, but are found on several localities, such as Øvre Kvam near Snåsavann.

På Munkholmen ved Trondheim fant jeg sommeren 1957 en magnetitforekomst. Forekomsten ligger på holmens vestside i noen nakne berg nedenfor den gamle festningsmuren, fig. 1.

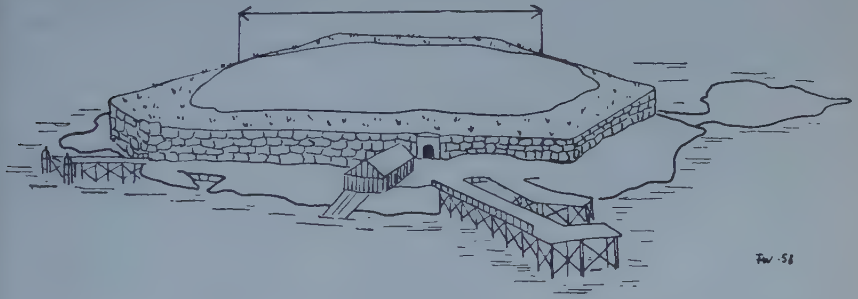

Fig. 1. Skisse av Munkholmen sett fra øst som viser forekomstens beliggenhet. Selve festningen er utelatt og en ser bare muren og vollene. Pilene angir forekomstens lokalitet.
Sketch of Munkholmen seen from the east. The arrows indicate the position of the locality depicted in Fig. 2.

En må gå så langt tilbake som til J. KRAFT (1832) for å finne forekomsten nevnt i litteraturen, og den kan i det hele ikke sees å være beskrevet. Jeg vil derfor beskrive den litt nærmere.

Lagserien på Munkholmen består av grønnstein med putestrukturer, blåkvartslag, ofte med kalkspat og magnetitrike lag som foruten magnetit inneholder både svovelkis og hornblende.

På fig. 2 sees i nummerrekkefølge ovenfra og nedover:

1. Lys grønnsten med putestrukturer. Putene er ca. 10 cm lange og et par cm brede.
2. Blåkvartslag. Mektighet ca. 20 cm.
3. Lys grønnsten med putestrukturer. Dimensjoner som ved 1.

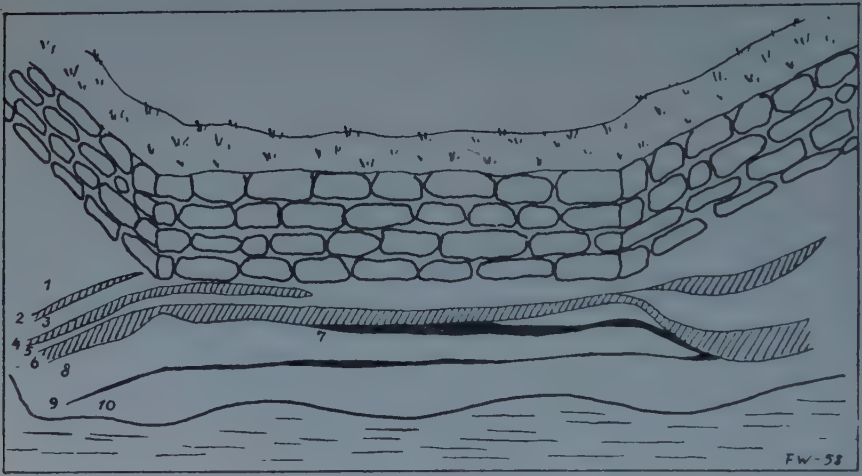

Fig. 2. Lagserien på Munkholmen sett på skrå ovenfra slik den trer frem mellom festningsmuren og sjøen på holmens vestside. De forskjellige lagene er nummerert fra 1 til 10 ovenfra og nedad.

Shore exposure at the west side of Munkholmen near Trondheim. 1. Light greenstone with pillow structure. The pillows are 10 cm long and about 2 cm thick. — 2. Layer of blue quartzite, thickness ca. 20 cm. — 3. Light greenstone with pillow structure. Dimensions of the pillows as at 1. — 4. Layer of blue quartzite, thickness ca. 40 cm. — 5. Light greenstone rich in $CaCO_3$*, with pillow structure. Dimensions of the pillows as at 1. — 6. Layer of blue quartzite with crack-fillings of* $CaCO_3$*, thickness from 1 m to 1.5 m. — 7. Layer rich in magnetite with hornblende and pyrite, thickness from 20 cm to 40 cm. — 8. Light greenstone rich in* $CaCO_3$*, thickness ca. 1 m. — 9. Layer rich in magnetite with hornblende and pyrite, thickness from 5 cm to 35 cm. — 10. Light greenstone rich in* $CaCO_3$.

4. Blåkvartslag. Mektighet ca. 40 cm.
5. Kalkrik, lys grønnsten med putestrukturer. Dimensjoner som ved 1. og 3.
6. Blåkvartslag med sprekkefyllinger av kalkspat. Mektighet fra 1 m til 1,5 m.
7. Magnetitrikt lag med hornblende og svovelkis. Mektighet fra 20 cm til 40 cm.
8. Lys kalkrik grønnsten. Mektighet ca. 1 m.
9. Magnetitrikt lag med hornblende og svovelkis. Mektighet fra 5 cm til 35 cm.
10. Lys kalkrik grønnskifer.

Blåkvartsen er ofte båndet i mørkere og lysere bånd. Dette peker hen på blåkvartsens sedimentære opprinnelse. Sprekker fylt med sekundært utfelt kalkspat og hvit kvarts skjærer båndingen.

Hornblendenålene i de magnetitrike lagene er av størrelsesorden 1 cm og opptrer i bånd. Rene magnetitlag opptrer bare sonevis. Ellers veksler de magnetitrike sonene med mer eller mindre hornblenderike soner. Noen steder sees lysere soner med epidot.

Lagenes fall er omkring 30° mot sydøst. I strøkretningen mot nordøst ligger Ladehammeren hvor TØRNEBOHM (1896) har nevnt en blåkvartsforekomst. Det er er derfor mulig at det er Munkholmens blåkvartssone som fortsetter over på Ladehammeren. Forøvrig nevner TØRNEBOHM (1896), TH. KJERULF (1871) og C. W. CARSTENS (1919) flere forekomster av blåkvarts i nærheten av Trondheim. I selve Trondheim by har jeg selv på flere steder sett mindre lag av blåkvarts. Disse blåkvartslagene kan på grunn av sin beliggenhet ikke høre til samme horisont som blåkvartslagene på Munkholmen. Det antas derfor at det er en rekke blåkvartslag i Støren — Bymarkgruppens grønne tuffer og lavaer og at blåkvartsen på Munkholmen er en av disse.

Magnetitrike lag assosiert til blåkvartslagene finnes på flere steder innenfor Støren—Bymarkgruppen, men en finner ofte blåkvartslag uten at det er utviklet noen magnetitrike lag samtidig. Slike magnetitrike lag assosiert til blåkvartslag finnes i Øvre Kvam ved Snåsavann (H. CARSTENS 1955 og 1956) og andre steder innenfor Støren—Bymarkgruppen. Forekomsten i Øvre Kvam hadde jeg anledning til å undersøke i detalj sommeren 1955 og den har mange likhetstrekk med forekomsten på Munkholmen.

Institutt for Geologi, Universitetet i Oslo, november 1958.

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Trykt mai 1959

Trondheims-feltet — Gneis-området — Jotunheims-området.

En kort bemerkning til Hernes' svar.

I sitt svar til mitt innlegg fremholder Hernes at han i sin avhandling «Connections —» først og fremst har behandlet den kaledonske fjellkjedes forløp i Sør-Norge, at den efter hans mening danner en bue. Han skriver at jeg øyensynlig har hatt i tankene om det vestlige gneisområde er «grunnfjell eller kaledonsk» og mener at det spørsmålet har lite med hans emne å gjøre. Det kan jeg ikke være enig i. Skal man finne ut en fjellkjedes forløp, om buet eller rettlinjet, må man først vite hvilke områder som stemmer overens med hensyn til bergarter og strukturer. Er dette på det rene, er det lett nok å trekke opp linjene. Når man som Hernes skal trekke en linje fra Trondheims-feltet og inn i Gneis-området, bør man ikke unnlate å nevne de forskere som først har utredet sammenhengen mellom områdene, i dette tilfelle Holtedahl og senere Gjelsvik. I all vitenskapelig diskusjon og forfatterskap er det en gyllen regel at den som først har publisert en iakttagelse eller oppdagelse, stor eller liten, også skal krediteres for den ved at hans navn og arbeide i det minste blir nevnt.

Hernes skriver at han har tatt sitt utgangspunkt i Th. Vogts «Revolusjonshistorie» fra 1928, han refererer Vogts fremstilling av den postmellomdevonske (svalbardiske) folding og hva Vogts resultater har å si for det spørsmål han behandler. Det er utmerket at Hernes i dette tilfelle har ydet rettferdighet til en av sine forgjengere. Men når det gjelder arbeider over bergarter og strukturer fra de kaledonske hovedfaser, har Hernes ikke funnet det nødvendig å nevne de forfattere som her har prioriteten.

Blindern, november 1958

Trygve Strand

NY LITTERATUR

Grundzüge der zoologischen Mikropaläontologie. Band I. Av VLADIMÍR POKORNÝ. 582 + XII sider, 459 bilder. VEB Deutscher Verlag der Wissenschaften. Berlin 1958. Pris DM 48.—.

Den vitenskapelige og økonomiske betydning av mikropaleontologi og mikrostratigrafi har forståelig nok ført til at det nå blir undervist i dem som selvstendige fag-grener ved universitetene i alle kulturland. — Slik står det i forordet til denne boken, og man kan da bare håpe at også Norge med tiden skal kunne regnes som et kulturland. Det er riktig nok at mikropaleontologien særlig har vært anvendt ved leting etter olje, noe som jo ikke har vært aktuelt i Norge. Men foruten botanisk mikropaleontologi er zoologisk mikropaleontologi av praktisk betydning også her i landet, som vist av de senere års undersøkelser ved N.G.U. over foraminiferer i våre kvartære leirer.

Det ville være ønskelig om det ble undervist mere i mikropaleontologi ved de norske universiteter. Pokornýs lærebok i zoologisk mikropaleontologi er, såvidt jeg vet, den mest utførlige i sitt slag og vil egne seg godt som kilde for en slik undervisning. Boken utkom først på tsjekkisk i 1954 og er blitt oversatt til tysk etter oppfordringer om å få den trykt på et verdenssprog. Det er forståelig at det er behov for gode lærebøker i dette emne; for tiden publiseres det jo mange flere arbeider i mikropaleontologi enn i de andre grener av paleontologien.

Foreløpig er bare første bind kommet ut på tysk. Det omfatter greie oversikter over mikropaleontologiske arbeidsmetoder og systematisk gjennomgåelse av foraminiferer og andre encellede mikrofossiler. Foruten systematikk (ned til underslekt) og detaljerte morfologiske beskrivelser, omtaler boken heldigvis også inngående hva man vet om de forskjellige gruppers økologi, forplantning, ontogenese og phylogenese. Såvidt jeg kan bedømme uten selv å ha arbeidet med disse grupper, virker det som om Pokornýs bok er grundig og up-todate. Foruten oversikter over de forskjellige teorier i spørsmål som blir behandlet, leverer forfatteren ofte egne bidrag. Det er tatt med en mengde stoff av mer generell interesse, f. eks. om utviklingsprinsipper og aktualitetsprinsippet. Det er en fordel at forfatteren er vel kjent med både «østlig» og «vestlig» litteratur, mens f. eks. amerikanske og russiske lærebøker ofte kan være ensidige. Boken inneholder et stort antall gode illustrasjoner. Av stor nytte er de ikke mindre enn 63 sider med litteratur-henvisninger, ordnet i 22 lister etter emne. Bokens annet bind vil behandle ostrakoder og andre mikrofossiler som hører til de flercellede dyregrupper.

For en som arbeider med fossiler fra kambro-siluren, er det verdt å notere seg det store antall protozoer som er beskrevet fra disse lag i andre land, både av radiolarer, foraminiferer, chitinozoer og hystricosphærer.

Paleontologisk Museum, Oslo.

Gunnar Henningsmoen

Précis de Pétrographie, Roches Sédimentaires, Métamorphiques, et Éruptives, av Jean Jung. 314 s. Masson et Cie, Paris-6e, 3.600 fr.

Dette er en utmerket bok, og meget vel egnet for den som vil studere petrografi og lære seg å kjenne de forskjellige bergarter i mikroskopet.

Ved hjelp av velvalgte og godt utførte tegninger illustreres de forskjellige bergartstyper med hensyn til struktur og mineralinnhold - der er tilsammen 200 figurer, derav 140 tegninger av tynnslip og 40 fotografier av håndstykker. På grunn av de mange figurer, er lesningen meget lett, og også av denne grunn anbefaler jeg boken til dem som vil forsøke seg på en fransk tekst.

Først gjennomgås de bergartsdannende mineraler (40 sider). Så følger de sedimentære bergarter (92 sider), metamorfe bergarter (55 sider), og til slutt eruptive bergarter (106 sider).

Skjønt hovedvekten legges på en beskrivelse av de mange bergartstyper, er også det teoretiske grunnlag behandlet, således at boken danner et vel avrundet hele. Jeg kan også være med når forfatteren uttaler at sammenlignet med sedimenter og metamorfiter er de eruptive bergarters verden på mange måter «beaucoup plus mystérieux».

Der er alltid enkelte små ting som kan bemerkes: Under melilit behandles ikke den Na-holdige komponent. Forskjellen mellom mikroklin og orthoklas ligger ikke i aksjevinkelens størrelse men i symmetriegenskapende (skjev, henholdsvis rett utslukning på basis, etc). Larvikit skrives Laurvickit.

Men dette er rene bagateller. Boken gir på en usedvanlig lettfattelig måte det meste av alt hva man kan lære av bergartsbestemmelse med mikroskopet.

Tom. F. W. Barth

NORSK GEOLOGISK FORENING

Møte nr. 378, 6. november 1958

på Institutt for geologi, Blindern.

Til stede: 30 medlemmer, 14 gjester.

Følgende innvalg ble godkjent:

430. Dr. Hans-Peter Geis, Vigsnes Kobberverk A/S, Vigsnes. Foreslått av Ferd P. Egeberg og F. Skjerlie.
431. Bergingeniør Geir Strand, Lalm. Foreslått av Chr. Oftedahl og A. O. Poulsen.
432. Bergingeniør Per Sandvik, Løkken Verk. Foreslått av T. Strand og K. Brøndbo.

Foredrag av:

Professor *O. Holtedahl:*
«Noen iakttagelser fra Grønsennknipa i Vestre Slidre.»

Professor *T. Strand:*
«Valdres-sparagmittens stratigrafiske stilling.»

I ordskiftet etter foredragene deltok:
B. Dietrichson, L. Størmer, Chr. Oftedahl, J. A. Dons, P. Holmsen og foredragsholderne.

Møte nr. 379 (julemøte) 11. desember 1958

i Drammensveien 78, Oslo.

Til stede: 32 medlemmer, 26 gjester.

Følgende innvalg ble godkjent:

433. Tegnesjef Olaf H. Halfdansen, Oslo. Foreslått av T. Strand og F. Hagemann.

Foredrag med fargelysbilder av:

Førstekonservator *Gunnar Henningsmoen:*
«Rundt omkring i Nord-Amerika.»

Generalforsamling torsdag 12. februar 1958

i Drammensveien 78, Oslo.

Årsmelding og regnskap for 1958 ble godkjent.

Årsmelding fra Bergens og Trondheims geologiske klubber ble referert. Fra professorene Barth, Heintz og Kolderup var det kommet følgende forslag til lovendring:

Nåværende § 5: Foreningens tidsskrift, Norsk Geologisk Tidsskrift, redigeres av styret med den valgte redaktør som hovedredaktør. Det utsendes såvidt mulig regelmessig med 4 hefter om året.

Forslag til ny § 5: Foreningens tidsskrift, Norsk Geologisk Tidsskrift, redigeres av den valgte redaktør og en valgt redaksjonskomité, som består av 3—5 medlemmer. Redaksjonskomitéens medlemmer bør velges slik at de sammen dekker et så stort felt av fagområdet som mulig. Manuskripter skal godkjennes til trykning eller refuseres av redaktøren og minst ett av redaksjonskomitéens medlemmer. Eventuelle uenigheter innen redaksjonen avgjøres av styret. Tidsskriftet utsendes såvidt mulig regelmessig med 4 hefter om året.

Utdrag av nåværende § 7: På generalforsamlingen velges styre samt 2 revisorer med varamenn for det kommende år.

Denne setning foreslås forandret til: På generalforsamlingen velges styre samt 2 revisorer med varamenn og redaksjonskomité for det kommende år.

Forslaget ble vedtatt.

Valg:

Formann: *T. Strand.*
Viseforman: *N. Spjeldnæs.*
Sekretær: *K. S. Heier.*
Redaktør: *N.-H. Kolderup.*
Redaksjonskomité: *T. Barth, J. Bugge, A. Heintz, O. Holtedahl.*
Styremedlemmer: *K. Egede Larssen, A. Eriksen* og formennene i Bergens og Trondheims geologiske klubber.
Reuschmedaljekomité: *A. O. Broch, Chr. Oftedahl, T. Strand.*
Revisorer: *T. Gjelsvik, T. Winsnes* med *G. Henningsmoen* og *T. Siggerud* som varamenn.

A. Granli ble anmodet om å forsette som foreningens forretningsfører.

Reuschmedaljen for 1958 ble tildelt statsgeolog, cand. real. Rolf W. Feyling-Hanssen for hans avhandling: «Late-pleistocene Foraminifera from the Oslofjord Area, Southeast Norway.» Norsk Geol. Tidsskr. Bd. 33, s. 109—152, 1954.

Årsmelding for 1958.

Siden forrige generalforsamling er følgende medlemmer avgått ved døden:

Professor H. Backlund
Bergingeniør A. Sunde
Professor Th. Vogt

2 medlemmer er utmeldt av foreningen.

I samme tidsrom er det innvalgt 14 nye medlemmer.

Foreningen har nå 251 medlemmer, derav:

2 korresponderende,
125 livsvarige,
124 årsbetalende.

Det er i siste år avholdt 6 ordinære møter med et samlet fremmøte av 252 personer.

Av Norsk Geologisk Tidskrift er utkommet Bd. 38, h. 1—4.

Utdrag av regnskap for det ordinære budsjett for 1958

Inntekt

Beholdning, overført fra 1957: I kasse	kr. 1 105,45		
I bank	» 3 474,36	kr.	4 579,81
Innbetalt kontingent		»	1 710,00
Tilskudd fra Staten		»	1 200,00
Tilskudd fra Norges Almenvitenskapelige forskn. råd . .		»	19 219,00
Ekstraordinært tilskudd fra Norges Almenvitenskapelige forskn. råd		»	13 786,00
Abonn. og salg av tidsskriftet		»	4 924,10
Renter av bankinnskudd		»	65,00
Renter fra livsvarige medlemmers fond		»	865,12
Renter fra Berg- og Steinindustriens fond		»	600,15
		kr.	46 949,18

Utgift

Tidsskriftet:			
Trykking Bind 37, hefte 3—4 og 38, hefte 1 og 2 . .		kr.	25 516,00
Trykking Bind 38, hefte 3—4 (avdrag)		»	6 373,00
Klisjeer Bind 37, hefte, 3—4 og 38, hefte 1—4		»	8 683,00
Korrektur og tegnearbeide		»	994,00
Ekspedisjon av tidsskriftet		»	1 600,50
Arbeidshjelp, porto og skrivesaker		»	664,55
Møter og ekskursjoner		»	500,36
Representasjoner, anskaffelser og Reuschmedaljen		»	35,16
Beholdning, overført til 1959: I kasse	kr. 504,13		
I bank	» 2 078,48	»	2 582,61
		kr.	46 949,18

Livsvarige medlemmers fond 1958.

		Inntekt	Utgift
1/1 1959. Fondets kapital: .	kr. 19 710,00		
Innbet. kont. i 1958 . . .	» 1 550,00		
	kr. 21 260,00		
Herav i statsobl. og aksjebrev	kr. 15 100,00		
I bank	» 6 160,00		
31/12 1958. Urørlig kapital	kr. 21 260,00		
Renter av verdipapirer for 1958		kr. 735,50	
Renter av bankinnskudd for 1958		kr. 152,62	
Forvaltning av verdipapirer			kr. 23,00
Overført til ordinære budsjett			» 865,12
		kr. 888,12	kr. 888,12

Berg- og steinindustriens gavefond.

		Inntekt	Utgift
Fondets kapital	kr. 20 000,00		
Renter av bankinnskudd		kr. 600,15	
Overført til ord. budsjett			kr. 600,15
		kr. 600,15	kr. 600,15

Regnskapene revidert av G. Henningsmoen og T. Siggerud 17/1 1959.

Møte nr. 380, 12. februar 1959

i Drammensveien 78, Oslo.

Til stede 42 medlemmer, 14 gjester.

Følgende innvalg ble godkjent:

434. Statsgeolog Trygve Eriksson, Sveriges Geologiska Undersökning, Stockholm. Foreslått av P. Padget og G. Kautsky.
435. Sjefsgeolog dr. Otto Meier, Svenska Diamantbergborrnings A/B, Hamngt. 29, Sundbyberg. Foreslått av Chr. Oftedahl og T. Siggerud.

Direktør *Carl Bugge* holdt minnetale over professor Thorolf Vogt.

Foredrag av statsgeolog *Chr. Oftedahl:*
«Om fjellkjedens vulkanitter».

I ordskiftet etter foredraget deltok:
I. Rosenqvist, A. Kvale, T. Strand og foredragsholderen.

Bergens Geologiske Klubb

Årsmelding for 1958.

Møte nr. 67, 5. februar.

32 tilhørere.

Professor dr. *Axel Sømme:* «Litt av hvert fra Japan.» Etter foredraget festlig sammenkomst på Geologisk institutt i anledning professor Niels-Henr. Kolderups 60-års dag.

Møte nr. 68, 6. mars.

20 tilhørere.

Dr. Ulf Hafsten: «Strandlinjeforskyvning i Oslo-trakten under senkvartær tid.»

Møte nr. 69, 27. mars.

18 tilhørere.

Professor dr. *Anders Kvale* og cand. real. *M. A. Sellevoll:* Orientering om jordskjelvstasjonen på Svalbard, med demonstrasjon av seismografen.

Møte nr. 70, 3. desember.

16 tilhørere.

Dr. *Hans Holtedahl:* «Geologiske og mikropaleontologiske undersøkelser av sedimenter fra Norskehavet.»

Møte nr. 71, 13. desember. Julemøte.

44 tilhørere.

Dr. *Tore Sund:* «Tid og rom i Skottland.» Etter foredraget selskapelig samvær på Geologisk institutt.

Styret var i 1958:

Formann:	Professor Anders Kvale,
Viseformann:	Lektor A. Samuelsen,
Sekretær:	Amanuensis Ivar Hernes.

GEOLOGNYTT

Universitetet i Oslo

Cand. real. EGIL SÆTHER ble av Det akademiske kollegium i juni 1958 kreert til doctor philosophiæ for sin avhandling "The Alkaline Rock Province of the Fen Area in Southern Norway". Det Kgl. Norske Videnskabers Selskabs Skrifter 1957 Nr. 1.

Norges geologiske undersøkelse

Vitenskapelig assistent KARI EGEDE LARSSEN er ansatt som statsgeolog II fra 1. oktober 1958.

Midlertidig statsgeolog II THOR LORCK SVERDRUP, er fast ansatt i stillingen fra 16. november 1958.

Sivilingeniør ROAR SOLLI er beskikket som laboratorie-ingeniør fra 1. februar 1959.

Cand. real. KNUT ØRN BRYN er ansatt som midlertidig vitenskapelig assistent fra 1. november 1958.

Universitetet i Bergen

Tidligere stipendiat ved Pollenanalytisk institutt, dr. philos. ULF HAFSTEN, er fra 1. januar 1959 beskikket som dosent i botanikk ved Botanisk museum.

Eksamener

Ved Universitetet i Oslo er i høstsemesteret 1958 matematisk-naturvitenskapelig embedseksamen med hovedfag mineralogi og petrografi fullført av:

AUDUN HJELLE, med hovedoppgave «En del undersøkelser over grunnfjellets petrografi, tektonikk og alminnelig geologi i deler av Stange og Romedal herreder, Hedmark fylke.»

OLE-KRISTOFFER IHLE, med hovedoppgave «Kvartærgeologien i Ringsaker og deler av Brøttum.»

I vårsemesteret 1959 av:

CHR. DICK THORKILDSEN, med hovedoppgave «Geologisk undersøkelse av bergartene i den østre del av anorthositt-området i Egersundfeltet, Rogaland fylke.»

EVOLUTION OF PTYGMATIC FOLDING

BY

HANS RAMBERG

(University of Chicago)

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Abstract: It is argued that ptygmatic veins generally are caused by a component of compressive strain parallel to the vein. Pinch-and-swell and boudinage structures develop when the veins are so oriented that a component of extensive strain is parallel to the veins.

Analysis of strain geometry give relationships between angular attitude of veins (relative to the principal strain axes in rocks) and magnitude of stretching (pinch-and-swell), or magnitude of shortening (ptygmatic folds).

When rocks are strained plastically there always are some directions along which compressive longitudinal strain occurs and others along which extensive longitudinal strain occurs. Ptygmatic folding and pinching-and-swelling or boudinage are consequently complimentary phenomena which can develop simultaneously in response to the same stress system in the host rock.

Experiments with putty as host and plasticene sheets as veins give results identical to natural ptygmatic and, pinch-and-swell structures.

The longitudinal compressive force which is responsible for buckling of veins is caused by drag or *skin friction* along veins due to plastic compressive flow in less competent host rocks. An equation for this compressive force as a function of dimensions of vein and principal stresses in the host rock is developed.

The buckling of an enclosed vein is resisted by two distinct forces: 1) the strength (viscosity) of the vein itself, and 2) the strength (viscosity) of the enclosing rock. Resistance 1) is at a minimum when the entire vein makes *one* half-wave; resistance 2) is least when the vein makes an infinite number of infinitely small waves. The actually developed wave length (or rather length of arc) represents a compromise between these two tendencies. Equations for the two resistance forces and the most "stable" initial wave length are presented. This initial wave length, λi, of ptygmatic veins is the same as the length of arc of more mature folds. λi is the wave length which requires minimum compressive force to develop.

It is implied that a study of geometry and dimensions of ptygmatic and pinch-and-swell veins in the field gives valuable information not only as to the geometric type of strain which has effected the considered rock complex, but also as to relative strength and creep viscosity of rocks.

Introduction.

When Sederholm in 1913 first coined the term ptygmatic folds for the sinuous shape of many veins in gneisses and migmatites, he considered the folding as a secondary phenomenon caused by shortening of plastic host rocks in directions parallel to the veins. This conclusion has in principle been accepted by several students of such veins (Erdmannsdörfeer, 1938; Kuenen, 1938; Spurr, 1923: Godfrey, 1954; Ramberg, 1952–56). Other geologists, however, have suggested that ptygmatic structure develops without host-rock compression. Read, 1928, assumes that emplacement along irregular zig-zag fracture systems results in ptygmatic veins, and Niggli, 1925, Wilson, 1952, proposed that forcefull injection of magma in incompetent rocks gives rise to ptygmatic structures. The writer has had opportunity to study numerous examples on veins with ptygmatic structure in gneisses and schists, mostly pegmatites and aplites in high grade gneisses and quartz-calcite veins in low grade schists, and he has found no reason to doubt that this striking structure essentially is caused by a component of shortening in host rocks parallel to originally more or less planar veins. Most arguments against this view are probably due to lack of obvious strain features in many ptygmatic veins. This fact, however, only shows that the deformation takes place under somewhat different conditions (greater plasticity or slower strain rate) than those existing under evolution of rocks commonly accepted as tectonites. Strain experiments with putty and plasticene subsequently to be discussed have brought convincing support to the strain theory of ptygmatic folds.

At the proper difference in competency between vein and host rock — the vein must be somewhat more competent than the host — the vein can only adjust to plastic compressive strain in the host rock by a folding mechanism, provided the vein is properly oriented in relations to the stain geometry of the host rock. In this connection one may note that Sederholm (1913 and 1926) seems not to have realized the necessity of the vein material being more competent (mechanically strong) than the host. In his first paper, in fact, he assumed that the vein material was liquid during the folding because of lack of catalastic phenomena in the vein. In his later paper, however, Sederholm considered the folding to have occurred in the crystalline state, the lack of cataclasis being explained by simultaneous recrystallization.

Plastic compression along some directions in rock complexes must be associated with simultaneous extensions in directions that make large angles to the compression, provided that volume remains essentially constant during strain. Relatively competent sheets of rocks which make large angles with direction of maximum compression will therefore develop tension fractures or necked-down regions. One should consequently expect boudinage and pinch-and-swell structures to be associated with ptygmatic structures in the field as unseparably as extensive strain is associated with compressive strain in plastic deformation at constant volume. Kuenen, 1938 p. 23 notes such field association between stretching and ptygmatic folds. Field experience of the present writer supports this expectation: pinch-and-swell structure of quartz-feldspar veins or boudinage structure of other competent sheet-shaped rocks are generally so intimately associated with ptygmatic structures in the field that a complementary relationship of the kind mentioned above is strongly suggested, see Ramberg, 1956, p. 192.

Boudinage and pinch-and-swell structures as consequense of plastic extensive flow in heterogeneous rocks have been discussed in some detail in a previous paper, (Ramberg, 1955). In the present paper emphasis is put on plastic shortening in strained rocks and the consequent evolution of ptygmatic veins.

Ptygmatic folds and pinch-and-swell structures in relation to geometry of strain.

Strain experiments with putty as incompetent host and plasticene as competent vein have resulted in structures strikingly similar to those seen in many veined schists and gneisses (migmatites), see pls. 1 to 6. The evolution of such structures is probably most rigorously explained in view of the general geometry of plastic strain. If the geometry of strain is studied in detail we find that many structural features, which have been considered inconsistent with the secondary-folding theory of ptygmatic veins (Wilson, 1952, p. 16), actually are to be expected when competent sheets are embedded at variuos angles in yielding incompetent matter. Such structural features supposed to be characteristic of primary injection of ptygmatic veins, also developed in the strain experiments performed by the writer, see p. 127 .

To clarify the relations between host-rock structure and the structure of deformed enclosed veins, we shall consider the geometry of strain. For simplicity we shall assume essentially homogeneous strain in the incompetent host rock (or putty) except for the unavoidable boundary effects close to ptygmatically folded or pinching-and-swelling veins. Let us consider the following geometric classes of homogeneous plastic strain:

1) Pure shear (in two dimensions): Maximum compressive strain parallel to z-axis in a rectangular coordinate system; maximum extensive strain parallel to x-axis, and zero strain along y-axis. No rotation of coordinate axes during strain; axes for principal stress coincide with axes of principal strain.
2) Irrotational three dimensional homogeneous strain:
    a): Maximum compressive strain parallel to z-axis; extensive strain equal in all directions in the x, y-plane. (Uniaxial compression.)
    b): Maximum extensive strain parallel to z; compressive strain equal in all directions in the x, y-plane. (Uniaxial extension.)
3) Simple shear or rotational homogeneous strain. Shearing strain by movement parallel to x, y-plane in x-direction. Length of dimensions along z, x and y remain constant.

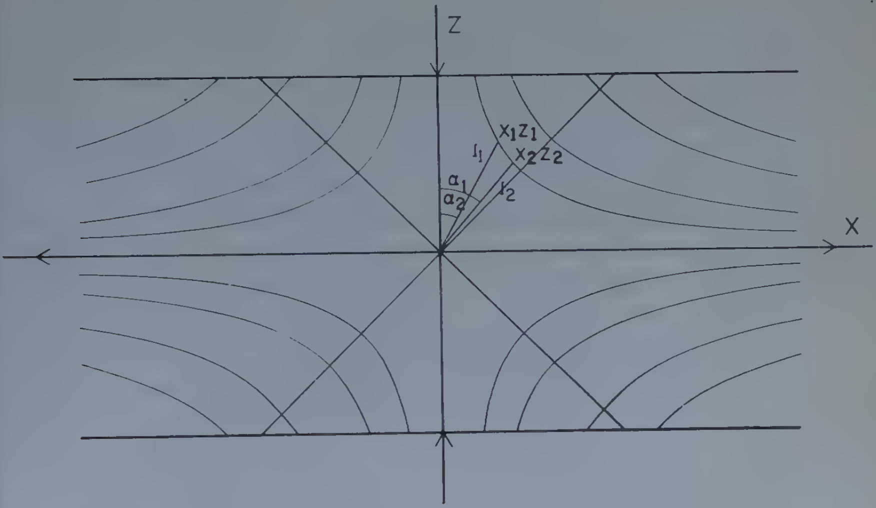

Fig. 1: Hyperbolic flow lines in a x, z section of a body under pure-shear deformation. The traces of the two 45° planes of no infinitisimal longitudinal strain are shown.

Pure shear deformation.

To make the strain homogeneous throughout the incompetent body, we must assume frictionless interfaces between the strained body and the medium which transmits stresses. This is not generally true in rocks, but the condition of homogeneous pure shear may be approached rather closely within a limited volume in the interior of a large homogeneous rock body.

Under pure strain the flow lines in the x, z-plane are families of rectangular hyperbolas with origin in the center of the strained body if we assume constant volume during strain, and disregard elastic effects, see fig. 1. During this kind of plastic strain, then, any given small particle in the body moves along a hyperbola of the form $x \cdot z = k$ where x and z are abscissa and ordinate respectively, and the constant k equals the area of the rectangle limited by the coordinates x and z and the coordinates axes. The hyperbolic shape of the flow lines follows from the requirment of constant volume during strain, and constant dimension of the strained body parallel to y.

Since the strain is homogeneous the movements in the environment to any given particle relative to that particle are the same independent of the position of the reference particle. The relative motion of matter in the vicinity of any given particle follows hyperbolic curves with center in the reference particle which itself of course is moving along a flow line. The relative flow is thus identical to the flow around the fixed origin of the coordinate system in the center of the plastic body. Any deformation which we now shall consider in relation to a coordinate system fixed in the center of the deformed body will consequently be identical to deformations in parallel directions anywhere throughout the strained body.

Since we are concerned with ptygmatic veins (plastic shortening of host rock parallel to vein), and pinch-and-swell structures of veins (streching parallel to vein) housed in plastic rocks, it is of prime interest to determine the angular attitude of directions along which compressive — respective extensive — strain occurs under pure shear. We shall firstly consider directions parallel to the x, z-plane.

Let x and z be coordinates to a point x, z, and l its distance from origin. We are seeking dl/l, i.e. the infinitisimal longitudinal strain along l, as a function of the angle between l and the z axis, and of the principal infinitisimal compressive strain in the body along z, dz/z.

According to fig. 1 the following relations are true:

1) $$l^2 = x^2 + z^2$$

2) $$x \cdot z = k.$$

Eq. (2) in (1) gives:

3) $$l^2 = \frac{k^2}{z^2} + z^2.$$

This can be differentiated:

$l \cdot dl = (z - k^2 z^{-3})\, dz$, which divided by eq. (3) gives the wanted quantity:

4) $$\frac{dl}{l} = \frac{1 - k^2 z^{-4}}{k^2 z^{-4} + 1} \frac{dz}{z}.$$

5) $$\tan \alpha = \frac{x}{z},$$

see fig. 1. Combining eq. (2) and (5) gives:

6) $$\tan \alpha = k z^{-2}.$$

Accordingly, eq. (4) can be written as follows:

7) $$\frac{dl}{l} = \frac{1 - \tan^2 \alpha}{1 + \tan^2 \alpha} \frac{dz}{z}.$$

The infinitisimal principal strain, $\frac{dz}{z}$ is negative in pure shear inasmuch as shortening takes place parallel to z. The sign of $\frac{dl}{l}$ which shows whether l shall suffer shortening or lengthening, depends then only upon the angle α as follows:

$$45^\circ < \alpha < 90^\circ \quad \text{gives} \quad \frac{dl}{l} > 0$$

$$\alpha = 45^\circ \quad \text{gives} \quad \frac{dl}{l} = 0$$

$$0 < \alpha < 45^\circ \quad \text{gives} \quad \frac{dl}{l} < 0$$

In words: at infinitisimal strain all directions making smaller angle than 45° with z become shortened by relative amounts varying from $\frac{dz}{z}$ at $\alpha = 0$ to zero at $\alpha = 45^\circ$. Directions making larger angle than 45° with z are lengthened by amounts varying from $\frac{dx}{x} = -\frac{dz}{z}$ at $\alpha = 90^\circ$ to zero at $\alpha = 45^\circ$, see fig. 1.

For directions deviating from the x, z-plane we find that infinitisimal lengthening and shortening respectively are confined to angular directions limited by two planes parallel to y and making 45° angle to z and x. Shortening effects directions in the angular region bisected by the y, z-plane, lengthening effects directions within the region bisected by the y, x-plane. The two sets of planes of no longitudinal infinitisimal strain which bisect the angle between the x- and z axes coincide with the circular cross sections of the strain ellipsoid at infinitisimal distortion from a sphere. However the angular attitude of the planes of no longitudinal infinitisimal strain remains constant at 45° inclination to z during the entire deformation whereas the circular cross sections in the strain ellipsoid of course rotate toward the x, y-plane.

During deformation matter moves at right angle across the planes of no infinitisimal longitudinal strain following the hyperbolic flow lines.

It is important for the following discussion to distinguish between imaginary mathematical planes or lines in the strained body, and

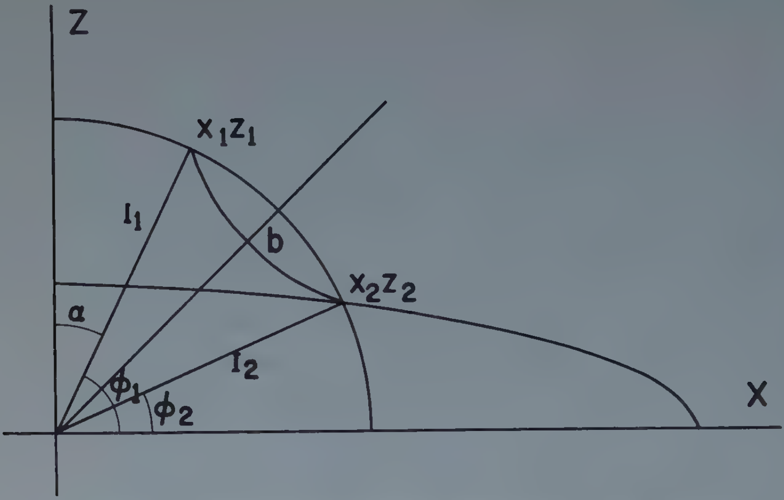

Fig. 2: Pure-shear type deformation of a sphere shown in one quadrant of the x, z-plane. 1_2 is the trace of the circular cross section. The curve x_1 z_1 -b-x_2z_2 is the hyperbolic trajectory of a particle existing at x_1 z_1 prior to straining and ending up at x_2z_2 after finite strain.

real sheets or rods which are made up of the constitutent particles of the yielding body or of some incorporated foreign material (veins, etc.). The movement and strain of such sheets and rods — their rotation, their longitudinal and shearing strain, their displacement in relation to each other etc. — are controlled by the movement of particles along the hyperbolic flow lines. One must bear in mind that the rotation and strain of materialistic sheets often is quite different from the rotation and strain of mathematical planes which can be constructed in the strain ellipsoid. The 45° planes of no infinitisimal longitudinal strain are imaginary mathematical sections through which matter flows. So are the circular cross sections in the strain ellipsoid; although these planes rotate during strain, properly oriented sheets of matter rotate faster and may sooner or later overtake the circular cross sections and rotate through them. The circular cross sections are planes of no finite longitudinal strains, i.e. $\frac{\Delta l}{l} = 0$, because any rod of matter from origin to the circumference of the circular cross section has the same length as this rod had in the original unstrained sphere. This does not mean, though, that rods or sheets lying parallel to the circular cross sections have not suffered longitudinal strain. Indeed such bodies have first been shortened and then stretched exactly to their original length during their rotation toward the circular cross sections in the strain ellipsoid, see fig. 2.

It is noteworthy that at the moment sheets or rods of matter rotate through the circular cross sections, the longitudinal infinitisimal strain, $\frac{dl}{l}$, as related to an infinitisimal angle of rotation, $d\alpha$, has a finite positive value. Only at the very instant the rotating sheets or rods are parallel to the 45° oblique planes of no longitudinal infinitisimal strain is $\frac{dl}{l} = 0$.

The above-mentioned difference between the circular cross sections in which finite longitudinal strain is zero and the 45° planes of no longitudinal infinitisimal strain is of consequence in the study of ptygmatic folding and pinch-and-swell structures.

The rotation of the circular cross section as a function of finite principal strain, $\frac{\Delta z}{z}$, follows from fig. 2.

We have:

8) $$\frac{z_2 - z_1}{z_1} = \frac{\Delta z}{z} = \frac{l_2 \sin \varphi_2 - l_1 \sin \varphi_1}{l_1 \sin \varphi_1}$$

where subscript 1 refers to original length and angle, subscript 2 to strained length and angle. Since we seek the angle of the circular cross section after finite strain, l_2 must be equal to l_1, and eq. (8) becomes:

9) $$\frac{\Delta z}{z} = \frac{\sin \varphi_2}{\sin \varphi_1} - 1.$$

l_1 and l_2 are symmetric in relation to the 45° plane of no infinitisimal strain, hence $\varphi_2 = \alpha_1$ or $\varphi_1 + \varphi_2 = 90°$ from wich follows: $\sin \varphi_1 = \cos \varphi_2$. Inserting this in eq. (9) gives

10) $$\tan \varphi_2 = \frac{\Delta z}{z} + 1\,,$$

where φ_2 is the angle between the x, y-plane and the circular cross section, and $\frac{\Delta z}{z}$ is finite principal strain in z direction.

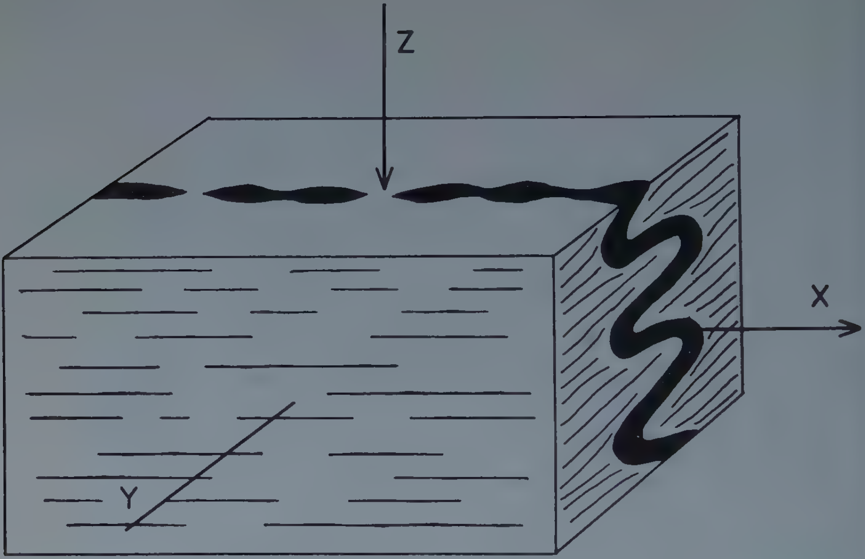

Fig. 3: Simultaneous development of ptygmatic folds and pinch-and-swell in a crosscutting vein under pure shear type deformation.

It is of interest for our purpose to determine interelations between angular attitude and angular rotation of embedded sheets (veins), longitudinal strain in various directions in such sheets, and strain of the whole incompetent rock body (putty) in the direction of principal strain axis.

Longitudinal strain in three sheets parallel to the coordinate planes is trivial. Sheets parallel to the x, z-plane are shortened by an amount $\frac{\Delta z}{z}$ parallel to z, and lengthened by an amount $\frac{\Delta x}{x}$, parallel to x. $\frac{\Delta z}{z}$ is finite compressive strain parallel to z, and $\frac{\Delta x}{x}$ finite extensive strain parallel to x.

If the enclosed sheet is identical to the surrounding body in mechanical properties or more soft, neither ptygmatic folding nor pinch-and-swell will develop as was verified by several tests. Sheets more competent than the host adjust to shortening in z direction by ptygmatic folding about an axis more or less parallel to x. In attempt to adjust to the extensive strain parallel to x the same sheet may develope pinch-and-swell (or boudinage) structure with necked down

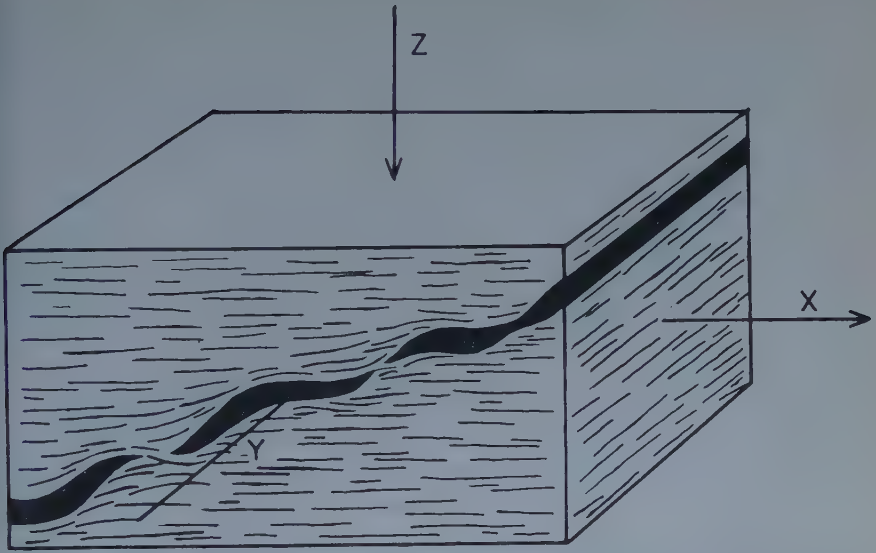

Fig. 4: Pinch-and-swell formed in a vein which was folded when oriented in the compression segment at the early stage of deformation, and stretched when rotated into the extension segment. The geometry of strain is pure-shear type.

zones parallel to the x, y-plane. Fig. 3 shows a block diagram of a vein deformed in this manner. It is worth noting at this point that pinch-and-swell (or boudinage) may develop by stretching along x in veins which cut across schistosity at a large angle. The kind of schistosity which is caused by strain in crystalline rocks is here assumed to develop perpendicular to maximum compressive finite strain, i.e. parallel to the x, y-plane in pure strain, rather than parallel to maximum shear strain.

Sheets parallel to the y, z-plane are shortened by an amount $\frac{\Delta z}{z}$ in the z direction and remain unstrained parallel to y. Ptygmatic folding about axis parallel to y is hence possible for competent veins oriented parallel to the y, z-plane in incompetent hosts, but pinch-and-swell structure cannot develop in sheets with this orientation.

Sheets parallel to the y, x-plane will be stretched in x directoin and unstrained in y. Such sheets may thus develop boudinage or pinch-and-swell by fractures or necked down regions parallel to y, fig. D, pl. 2. Ptygmatic folding is not possible.

The behavior of sheets inclined to some or all three co-ordinate axes is more interesting. Sheets parallel to y but oblique to z and x are effected by rotation and shortening or lengthening in the x, y-direction depending upon original and final angular attitude. Let α_1 and α_2 be original and final angles respectively, between z and the sheet. At $\alpha_1 < 45°$ shortening in the x, z-direction occurs until rotation reaches the 45° plane of no infinitisimal strain. In this period of deformation, then, ptygmatic folding may develope with axis parallel to y. Continued compressive strain along z, however, makes the ptygmatically folded sheet rotate past the 45° angle into the region of lengthening. As a result the previously formed folds may either be stretched out, as in some of the experiments, or the folded sheet may disrupt into a row of sigmoidal boudins (see fig. 4). Incidentally, if ptygmatic folding and subsequent stretching which result from strain and rotation are reversible phenomena there are some simple relationships between the original angular attitude of an unstrained vein (sheet), and the angle at which the ptygmatic folds have just been straightened to restore the original length. At the moment of complete straightening of the vein, it is parallel to the circular cross section in a strain ellipsoid formed from a sphere which was undeformed at the moment compressive strain and ptygmatic folding of the vein started. Under conditions of pure shear we know that sheets of matter being parallel to the circular cross sections have rotated through an angle twice the angular difference between the circular cross section and the 45° planes of no infinitisimal strain, see fig. 2. Hence if the angle between a straightened out vein and the plane of schistosity is φ_2 then the angle between the original underformed vein and the plane of schistosity was:

11) $$\varphi_1 = 90 - \varphi_2.$$

There also exists a simple relation between finite principal strain parallel to z, $\frac{\Delta z}{z}$, and the angular attitude of a vein at the moment it is straightened out to its original length. At this stage the vein lies parallel to the circular cross section in the strain ellipsoid, and eq. (10) p. 107. is valid:

12) $$\frac{\Delta z}{z} = \tan \varphi_2 - 1$$

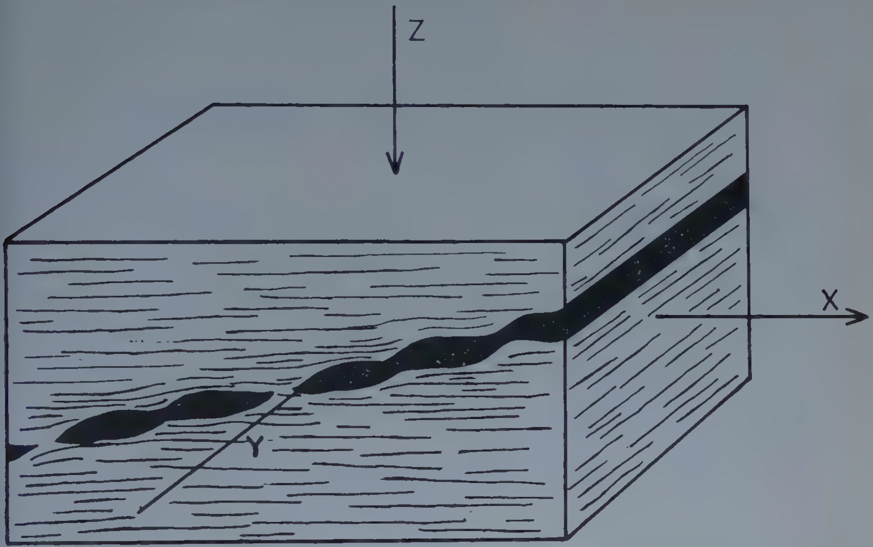

Fig. 5: Pinch-and-swell formed in a vein lying in the extension segment but yet inclined to the principal extension axis, x. Pure-shear type deformation.

where φ_2 is the angle between the straightened out vein and the plane of schistosity.

Eqs. (11) and (12) above are only rigorously applicable to pure shear strain in rocks if the deformations in the vein — its folding and subsequent straightening — are reversible and there is no slip between vein and adjacent rock. Such reversibility is probably not approached very closely in rocks. Some of the experiments, however, showed considerable reversibility of this kind, see for example fig. G, pl. 4.

Sheets parallel to y, but originally making greater than 45° angle with the z axis are effected only by stretching in the x, z-direction and may consequently develope pinch-and-swell or boudinage structure. It is interesting to realize that these structures thus can form in competent bodies with their long axis inclined to the schistosity in the host rock, see fig. 5.

Finite longitudinal strain of a sheet in the x, z-direction, $\frac{\Delta l}{l}$, is related to the angle of rotation, $\Delta\alpha = \alpha_2 - \alpha_1$ which in turn is related to the principal compressive strain of the whole plastic body,

$\frac{\Delta z}{z}$. These relationships may be useful as structural tools for if amount of stretching and shortening of a folded or pinching-and-swelling vein can be measured in the field, as well as its angle with the host-rock schistosity, the quantities $\Delta l/l$ and α_2 are known. α_1 and principal compressive strain along z in the host rock can then be estimated from the equations. (It is assumed here that the schistosity is produced by the same strain as that which produces ptygmatic folds and pinch-and-swell structures. If the observed schistosity is result of pre-vein deformation or sedimentary laying, or if it is parallel to shear strain rather than perpendicular to compressive strain, the following calculations do not hold.)

According to fig. 1 we can write:

13) $$l = \frac{z}{\cos\alpha},$$

hence:

14) $$\frac{\Delta l}{l} = \frac{l_2 - l_1}{l_1} = \frac{z_2\cos\alpha_1}{z_1\cos\alpha_2} - 1$$

Now $x \cdot z = k$, and $\tan\alpha = \frac{x}{z}$, therefore:

15) $$z = \sqrt{\frac{k}{\tan\alpha}},$$

which can be inserted in eq. (14):

16) $$\frac{\Delta l}{l} = \sqrt{\frac{\tan\alpha_1\cos\alpha_1}{\tan\alpha_2\cos\alpha_2}} - 1 = \sqrt{\frac{\sin\alpha_1\cos\alpha_1}{\sin\alpha_2\cos\alpha_2}} - 1$$

The angle α_1 can be found from eq. (16) if $\frac{\Delta l}{l}$ and α_2 are measured on ptygmatic — or pinch-and-swell veins in schists or gneisses. The principal compressive strain, $\frac{\Delta z}{z}$, can be estimated from the relation:

17) $$\frac{\Delta z}{z} = \sqrt{\frac{\tan\alpha_1}{\tan\alpha_2}} - 1,$$

which follows from eq. (15).

As an example on amount of longitudinal strain, $\frac{\Delta l}{l}$, caused by rotation of a vein from α_1 to α_2, we shall assume $\alpha_1 = 20°$ and $\alpha_2 =$

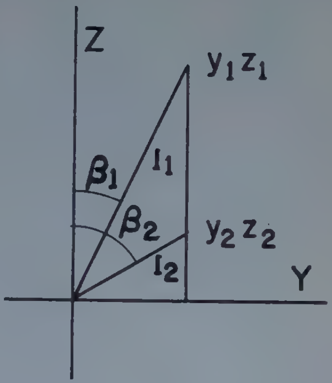

Fig. 6: Deformation of pure-shear type shown in one quadrant of a y, z section. Compression from z_1 to z_2 has caused shortening of line l from l_1 to l_2 and rotation from β_1 to β_2.

45°. Inserting these values in eq. (16) gives $\frac{\Delta l}{l} = -0.1925$ or 19.25 % shortening. Rotation from $\alpha_1 = 45°$ to $\alpha_2 = 80°$ results in $\frac{\Delta l}{l}$ 0.718, or 71.8 % lengthening.

During pure shear a sheet parallel to x but inclined to z and y will be stretched parallel to x by an amount $\frac{\Delta x}{x}$ and can consequently develop boudinage or pinch-and-swell structure with fractures or necked-down zones parallel to the y, z-plane. The y, z-direction in the sheet becomes shortened and rotates around the x axis which also is the axis of ptygmatic folds in the sheet. The amount of shortening in the y, z-direction, $\frac{\Delta l}{l}$, depends upon the angle, β, which the sheet makes with the z-axis before and after straining.

According to fig. 6 the following relation is valid:

$$l = \frac{y}{\sin \beta}, \tag{18}$$

where y is constant because in pure shear the motion of any given particle is parallel to the x, z-plane. The relative shortening in the y, z-direction in a vein is then:

$$\frac{l_2 - l_1}{l_1} = \frac{\Delta l}{l} = \frac{\sin \beta_1}{\sin \beta_2} - 1 \tag{19}$$

According to these equations one may for example estimate the principal strain in z of the host rock if $\frac{\Delta l}{l}$ and β_2 are measured on ptygmatic veins in the field. From fig. 6 follows equation

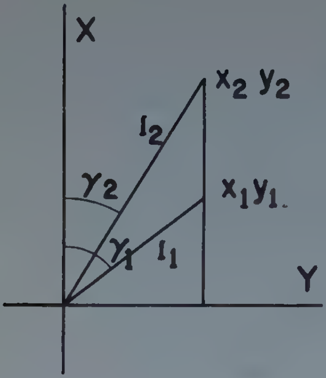

Fig. 7: Deformation of pure-shear type shown in one quadrant of the x, y-plane. Extension from x_1 to x_2 has caused elongation of line l from l_1 to l_2 and rotation from γ_1 to γ_2.

20) $$\frac{\Delta z}{z} = \frac{z_2 - z_1}{z_1} = \frac{\cot \beta_2}{\cot \beta_1} - 1$$

where β_2 is directly measurable and β_1 may be calculated from measured elongation, $\frac{\Delta l}{l}$, according to eq. (19).

Sheets parallel to the z axis but inclined to x and y are effected by principal compressive rock-strain parallel to z, and by various amount of stretching in x, y-direction depending upon the angle γ between the sheet and the x axis. Consider the strain in the x, y-plane in which the projected flow lines are parallel to the x axis, see fig. 7. According to this figure equation

21) $$l = \frac{y}{\sin \gamma}$$

is valid, where y is constant on account of the flow lines being parallel to the x, z-plane. Consequently the longitudinal strain is:

22) $$\frac{l_2 - l_1}{l_1} = \frac{\sin \gamma_1}{\sin \gamma_2} - 1$$

Measurement of elongation, $\frac{\Delta l}{l}$, (from separation of boudins or stretching of pinch-and-swell veins), and γ_2 enables one to calculate γ_1 by means of eq. (22). This is sufficient information, then, to estimate the principal extensive strain along x in the host-rock body according to equation: $\frac{\Delta x}{x} = \frac{\tan \gamma_1}{\tan \gamma_2} - 1.$

As the last example on sheet orientation under pure shear we shall consider sheets inclined to all three coordinates axes. It is convenient to let the sheet go through origin of the coordinate system. Any

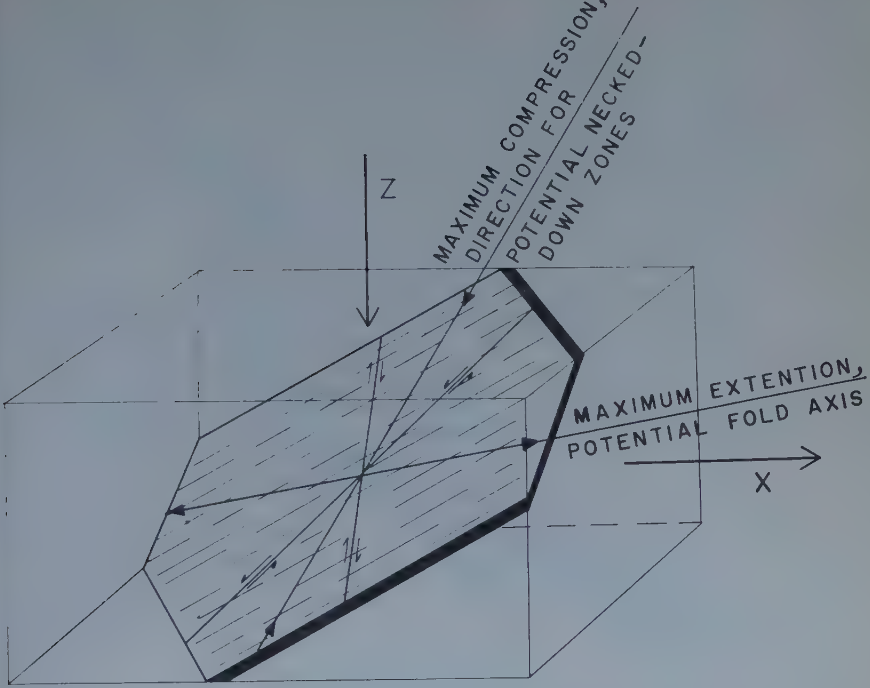

Fig. 8a: Block diagram of pure-shear effecting an oblique competent sheet of matter embedded in incompetent material. The directions of maximum compressive — and maximum extensive-strain in the sheet are shown. The directions of maximum shear and zero longitudinal strain are also indicated. These directions are the intersections between the oblique sheet and the two 45° planes of no longitudinal strain (not shown here).

oblique sheet through origin must intersect the three axial planes. Let us call the lines of intersection with the x, y-plane, the x, z-plane, and the y, z plane l_{xy}, l_{xz}, and l_{yz} respectively. The oblique plane is determined if two of the intersection lines are known. During pure shear of the host rock (putty) an embedded oblique sheet will both rotate and become effected by longitudinal strain. Rotation of the oblique sheet can be studied in terms of rotation of its lines of intersection with the axial planes. Relationships between principle rock strain in z or x, and increments in the angles α, β, γ are identical to those discussed above in connection with the three sets of sheets parallel to the y-, x- and z-axes respectively.

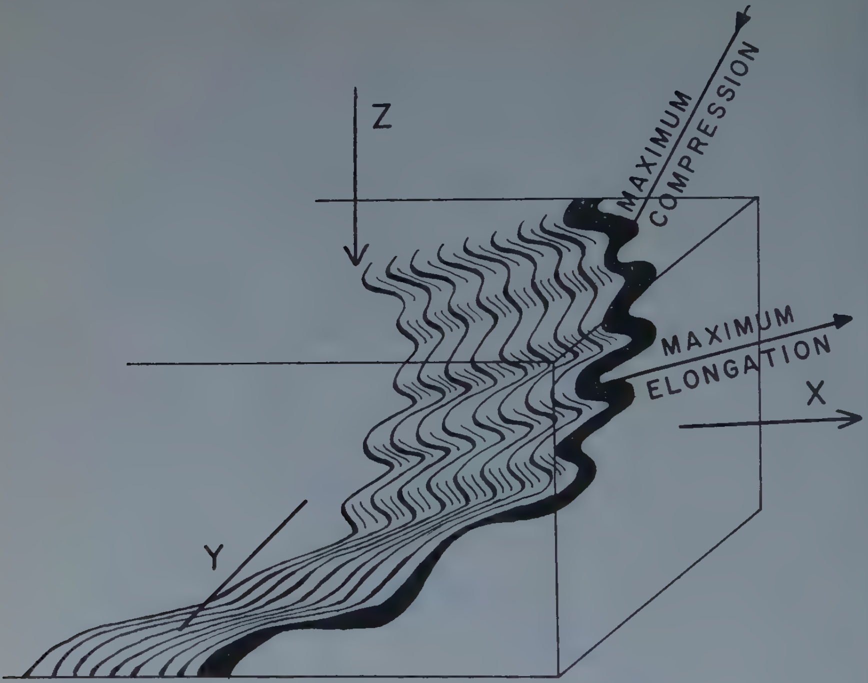

Fig. 8b: Oblique folding and stretching of the sheet shown in fig. 4a.

Any oblique sheet through origin must intersect the two 45° planes of no infinitisimal longitudinal strain. The angle between the two lines of intersection is generally oblique. These are directions of maximum shear and no infinitisimal longitudinal strain in the sheet. The angles between the two directions of no infinitisimal longitudinal strain in the oblique sheet are bisected by the directions of maximum positive and maximum negative longitudinal strain respectively. The direction of maximum positive longitudinal strain in the sheet lies closest to, but is not parallel to, the y, x-plane, and the direction of maximum negative longitudinal strain in the sheet lies closest to the z, y-plane, see fig. 8a, b. The intersections between the circular cross sections in the strain ellipsoid and the oblique sheet are lines of no finite longitudinal strain in that sheet.

In general, then, an embedded competent sheet oblique to all the principal strain axes is compressed into ptygmatic folds in the direction of maximum compressive longitudinal strain in that sheet and stretched into boudinage or pinch-and-swell structure in the direction of maximum extensive longitudinal strain. The axis of ptygmatic folds and the zones of necking down or boudin fractures of such oblique veins do in general make oblique angles with the host-rock schistosity and lineation. This conclusion is of significance in the argument as to whether ptygmatic folds are tectonic features or produced by primary injection.

Compressive strain parallel to z, uniform extensive strain in the x, y-plane.

The movement of any given particle in the yielding body follows a flowline lying in a radial plane which contains the z-axis. The equation for the flowline in the radial plane is:

23) $r^2z = k$ where $r = \sqrt{x^2 + y^2}$ is the distance from z-axis.

Flowlines in the x, z- and y, z-planes respectively, are given by eq. (24):

24)
$$\text{a) } x^2 z = k, \quad \text{b) } y^2 z = k.$$

Eqs. (23) and (24) follow from the condition of constant volume during plastic strain.

The family of flow surfaces are hyperbolic revolution surfaces rotated about the z-axis. We are interested in finding how longitudinal strain in an embedded sheet varies with its angular deviation from the z axis, and with the principal compressive strain in z. Because of rotational symmetry around z, all sections parallel to z give identical pictures with respect to the plastic strain. Let x be the abscissa, and z the ordinate in such a section. The distance from origin to a point x, z, is (fig. 9).

25)
$$l^2 = x^2 + z^2.$$

We are concerned about the infinitisimal longitudinal strain, $\frac{dl}{l}$, experienced by a sheet of matter making an angle, α, with z when it rotates through $d\alpha$, and the infinitisimal principal strain along z is $\frac{dz}{z}$.

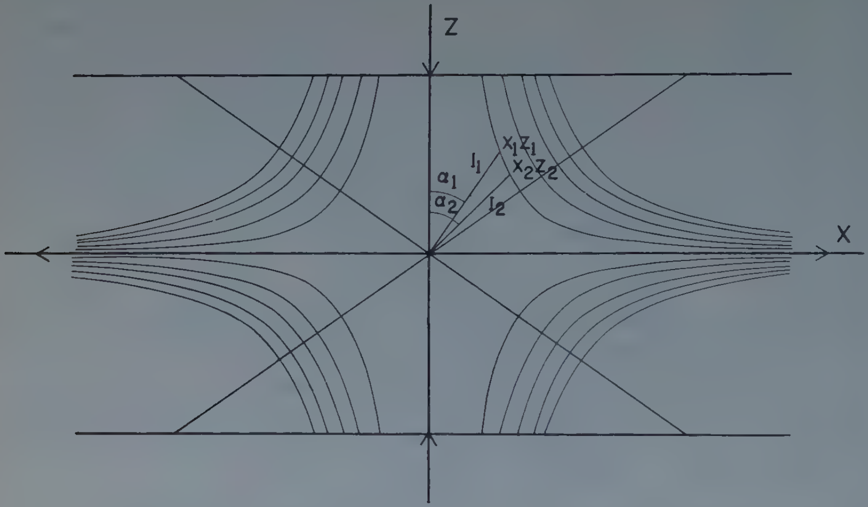

Fig. 9: Hyperbolic flow lines in a x, z section of a body under uniaxial compression. The traces of the two 54° 45′ planes of no inifinitisimal longitudinal strain are shown.

Eq. (24a) in eq. (25) gives:

26) $$l^2 = \frac{k}{z} + z^2.$$

Differentiating eq. (26) and dividing by itself results in:

27) $$\frac{dl}{l} = \frac{1 - 1/2\, kz^{-3}}{1 + 1/2\, kz^{-3}} \cdot \frac{dz}{z}.$$

28) $$\tan \alpha = \frac{x}{z} = \sqrt{kz^{-3}},$$

according to eq. (24). Consequently eq. (27) can be written:

29) $$\frac{dl}{l} = \frac{1 - 1/2 \tan^2 \alpha}{1 + 1/2 \tan^2 \alpha} \cdot \frac{dz}{z}.$$

Since the principal strain along z is compressive, $\frac{dz}{z}$ is negative and dl/l is zero when $1 - \frac{1}{2} \tan^2 \alpha = 0$. I.e. $\tan \alpha = \sqrt{2}$, and $\alpha = 54° 44'$. Thus for linear compression along z with uniform extension in all directions in planes perpendicular to z the directions with zero longitudinal infinitisimal strain lie on a double-cone surface with half angle equal to 54° 44′, and axis of revolution coinciding with the z axis. Directions with $\alpha < 54° 44'$ have negative dl/l and suffer shortening; directions with $\alpha > 54° 44'$ have positive dl/l and become stretched.

It is of interest to determine relations between angular rotation away from the z axis of a sheet of matter, $\Delta\,\alpha = \alpha_2 - \alpha_1$, and finite principal compressive strain in the host rock, $\Delta z/z$. According to eq. (28) z_1 and z_2 are related to α as follows:

$$\text{30)} \qquad \frac{\tan \alpha_2 - \tan \alpha_1}{\tan \alpha_1} = \frac{\sqrt{z_2^{-3}} - \sqrt{z_1^{-3}}}{\sqrt{z_1^{-3}}},$$

$$\text{31)} \qquad \frac{\tan \alpha_2}{\tan \alpha_1} = \sqrt{\frac{z_1^3}{z_2^3}}$$

Incidentally, the analogous relation for pure shear is:

$$\text{32)} \qquad \frac{\tan \alpha_2}{\tan \alpha_1} = \frac{z_1^2}{z_2^2}$$

from eq. (15) p. 112 showing that rotation of inclined sheets is somewhat larger for given principal compressive strain in pure shear than in uniaxial compression with rotational symmetry.

Finite longitudinal strain in the inclined sheet, $\frac{\Delta l}{l}$, as a function of rotation from α_1, to α_2 due to compression along z can be shown to be:

$$\text{33)} \qquad \frac{\Delta l}{l} = \sqrt[3]{\frac{\tan^2 \alpha_1}{\tan^2 \alpha_2} \frac{\cos \alpha_1}{\cos \alpha_2}} - 1$$

(Compare the corresponding equation for pure shear, eq. (16) p. 112).

Let us now consider the bahavior of competent sheets with various angular attitudes to the principal compressive strain axis. Sheets parallel to z will behave identically independent of their inclination to the x and y axis because of rotational symmetry of strain. Such sheets are shortened along z by an amount equal to the principal compressive strain in the host, and stretched in directions parallel to the x, y-plane. Ptygmatic folding and pinch-and-swell structures in corresponding directions are accordingly to be expected in analogy to sheets parallel to the x, z-plane in pure shear. However, in the present case the extensive strain in the x, y-plane is numerically smaller than the compressive strain in z.

Sheets perpendicular to z are stretched uniformly in all directions during strain thus giving rise to symmetric two dimensional boudinage or pinch-and-swell structures such as that shown in fig. C, pl. 6 in Ramberg, 1955.

All sheets inclined to z are stretched in directions parallel to the x, y-plane and may consequently develop boudinage and/or pinch-and-swell structures in that direction. As to whether an oblique sheet should become compressed or stretched in directions at right angle to its intersection with the x, y-plane depends upon the angle, a, between the sheet and z-axis. If $a > 54° 44'$, the said direction suffers stretching, and sheets with this inclination may develop two-dimensional pinch-and-swell structures. If $a < 54° 44'$ the same direction in the sheet is effected by shortening with consequent formation of ptygmatic folds.

The intersections between an oblique sheet and the cone-shaped surface of zero longitudinal infinitisimal strain correspond to directions in the sheet along which infinitisimal longitudinal strain vanishes.

Equal compression in all directions in x, y-plane, extension in z.

In a x, z- (or y, z-)plane the flow lines in this case also follow hyperbola of the character: $x^2 z = k$ (or $y^2 z = k$), and the flow surfaces are hyperbolic revolution surfaces with z as axis of rotation.

In order to determine the angular attitude of directions of no infinitisimal longitudinal strain as well as regions for compressive — and extensive strain respectively, we proceed as in the previous situation see eqs. (25) to (29) p. 118). If a is the angle between z and a given direction we find that infinitisimal longitudinal strain along this direction vanishes at $a = 54° 44'$. At $a > 54° 45'$ strain is compressive, i.e. $\frac{dl}{l} < 0$; at $a < 54° 44'$ strain is extensive i.e. $\frac{dl}{l} > 0$. The cone of zero infinitisimal longitudinal strain is thus the same as in uniaxial compression, but the regions of compression and extension respectively, have been reversed as compared with the previous case.

In analogy with the case of compression along z, it is simple to predict how embedded competent sheets with various angular attitude should behave under this type of plastic strain. A new situation however, occurs when sheets are more or less parallel to the x, y-

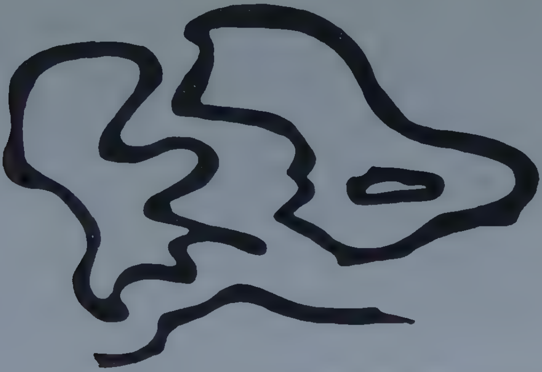

Fig. 10: A x, y section through a ptygmatic vein deformed in uniaxial extension type strain. The vein was originally a planar body oriented approximately parallel to the x, y plane.

plane which is effected by compressive strain in all directions. Such sheets must buckle in an intricate manner in attempt to adjust to shortening in all directions. Fig. B, pl. 6 shows experimental results of this kind of deformation. It is impossible to asign a fold axis to such crumpled veins, even if the host rock have developed a distinct fold axis and elongation during the same period of strain.

It is interesting to note that cuts across z through such complexly folded ptygmatic structures may show contorted veins with closed paths, see fig. 10. Cuts parallel to z, however, will reveal rather ordinary ptygmatic pattern. Under two dimensional compression two mutually perpendicular sheets both parallel to z will be ptygmatically folded simultaneously. Such features are not uncommon in veined gneisses and migmatites.

Simple shear deformation.

Let the movement be parallel to the x, y-plane and in the x direction. The two sets of planes in which longitudinal infinitisimal strain vanises, are parallel to the x, y-plane and the y, z-plane respectively. The direction of maximum infinitisimal extensive strain is parallel to the x, z plane and makes 45° angle with x; the direction of maximum infinitisimal compressive strain is parallel to the same plane and makes 90° angle with maximum extensive strain. Conse-

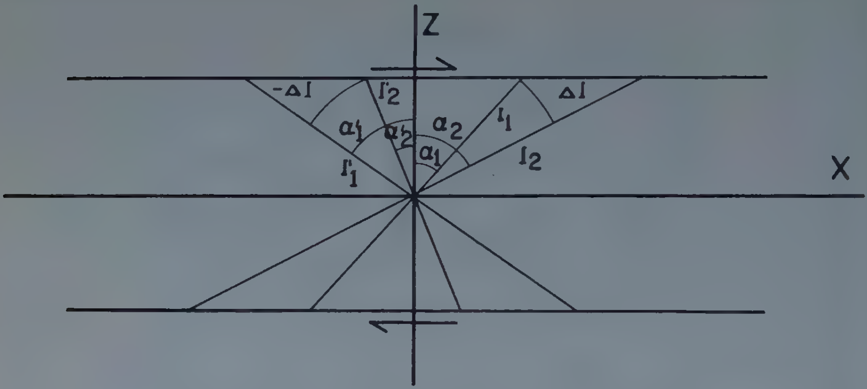

Fig. 11: Some geometric relationships in simple shear deformation.

quently all directions in the first and third quadrants are effected by infinitisimal extensive strain (positive $\frac{dl}{l}$), and directions in the second and fourth quadrant are effected by infinitisimal compressive strain, (negative $\frac{dl}{l}$).

The amount of finite longitudinal stretching or shortening of sheets as a function of their angular rotation during finite strain is readily found, see fig. 11:

34) $$l = z/\cos\alpha, \text{ and } \frac{\Delta l}{l} = \frac{\cos\alpha_1}{\cos\alpha_2} - 1,$$

where z is constant during simple shear.

$\Delta l/l$ is positive when $\alpha_2 > \alpha_1$ (only acute angles are considered). This can only happen either when the sheet originally is in the first and third quadrants or when a sheet in the second or fourth quadrants is rotated into the first and third quadrants through and angle of rotation $\Delta\,\alpha > 2\alpha_1$. If a sheet in the second quadrant is rotated less than $2\,\alpha_1$, then $\alpha_2 < \alpha_1$ and consequently $\frac{\Delta l}{l}$ is negative. The finite longitudinal strain caused by rotation from second to first quadrant is the result of a compressive strain (until $\alpha_2 = 0$) followed by an extensive strain. If $|\alpha_2| = |\alpha_1|$ or $\Delta\,\alpha = 2\,\alpha_1$, the compressive strain is exactly matched by the subsequent stretching and $\Delta l/l = 0$.

Whereas the one plane of no finite longitudinal strain rotates during deformation, both planes of no infinitisimal strain remain fixed parallel to the x, y- and y, z-planes during the entire deformation.

Let us now consider the theoretically expected behavior under simple shear of competent sheets embedded at various orientations relative to the coordinate axes. Competent sheets parallel to the x, y-plane should remain parallel to this plane. Since $\frac{dl}{l} = 0$ as well as $\Delta l/l = 0$ in this plane such sheets should neither suffer stretching nor folding, as was verified by experiments, figs. A, B, C, pl. 1.

Sheets parallel to x, z should be compressed and thrown into folds along the direction of maximum compressive strain, and stretched to pinch-and-swell or boudinage in the direction of maximum extension strain.

Sheets parallel to y but inclined to x and z are rotated and compressed to ptygmatic folds and/or stretched depending upon their original attitude and amount of shear movement, (pls. 1 to 6). The expected strain patterns of sheets inclined to all axes should be similar to those of sheets with corresponding attitude under pure strain as discussed above, fig. D, pl. 6.

Several experimental results of simple shear deformation are discussed below.

Experiments.

Putty was found to be a suitable imitation material for the incompetent host rocks, and plasticene, which is mechanically somewhat stronger than the putty, was used to simulate the veins. Incidentally, the same materials were used in the boudinage experiments of the writer (Ramberg, 1955). As to be expected, neither ptygmatic folding nor pinch-and-swell or boudinage developed under homogeneous strain when the «vein» material was identical to the host with respect to rheological properties or if the vein was softer as, for example, when colored putty was embedded in gray putty. Such embedded sheets remained straight during deformation. Longitudinal strain caused only thickening or thinning of the sheets. Of course, if deformation of the whole sample is non-homogeneous, as it tends to be close to boundaries, straight elongate inclusions of colored putty in grey putty would curve in various ways. This kind of deformation is outside the scope of this paper.

It was not practicable to make strain tests of all the possible orientations sheets discussed in the sections above, but sufficient number of runs were made to show that ptygmatic folding and pinch-and-swell or boudinage developed in excellent agreement with theoretical expectations.

Simple shear experiments.

It is relatively easy to maintain conditions of homogeneous strain in simple shear experiments. For this reason most experimental investigations of deformation of embedded sheets were made with this strain geometry. This fact, however, does not imply that the writer considers simple shear as the most common type of rock deformation. In his view the most frequent geometry of rock strain is one which contains an element of rotation but with all three principal strains unlike in magnitude and different from zero. Although the strain is rarely homogeneous over a large volume of rock complexes, it will essentially be so if the considered volume is small enough.

Two wooden blocks with a body of putty between were slid in shear movement relative to each other. The distance between the blocks was kept constant by placing a thin board of desired width between the blocks underneath the putty. The strain was homogeneous except in a narrow zone close to the wooden blocks and in the vicinity of the deformed sheets of plasticene. Because of friction between putty and the bottom board it is probable that the strain was non-homogeneous in the lower part of the putty body, but this did not effect the plasticene sheets or strips which were embedded in the upper part of the putty close to its free surface. In agreement with the theoretical models discussed in the previous section, the y-axis of the reference coordinate system is perpendicular to the free putty surface, the x-axis is parallel to the sliding movement, and the z-axis is perpendicular to the interfaces between putty and the wooden blocks. The plasticene sheets or strips, which were straight and uniformly thick, were embedded in the putty at various angles to the coordinate axes. Most sheets were so embedded that they did not touch the wooden blocks. Stress was therefore not transmitted to the ends of the plasticene sheets directly from the wooden blocks, but by frictional drag along the surfaces of the sheets caused by yielding putty. This is probably the most common situation in rocks also.

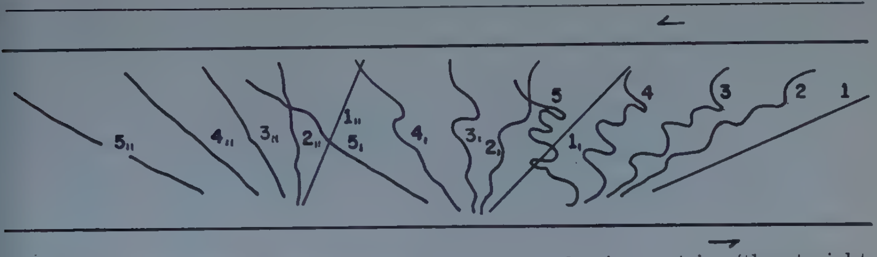

Fig. 12: Simple-shear type deformation of three plasticene strips (the straight lines marked by number 1) embedded in putty. 1, 2, 3, 4, 5 represents 5 steps of deformation. Traced after experiments.

Sheets embedded parallel to the x, y-plane remaine straight and unstretched during strain after even very large shear angle, see figs. A, B, C, pl. 1. This is to be expected in view of the fact that the x, y-plane is a plane of maximum shear strain and zero longitudinal strain.

Incidentally, the experimental fact that x, y-layers remain straight in simple shear suggests that dragfolds in layered rocks are not produced by shearing alone parallel to the folded layer; a component of compressive strain in direction along the layer seems imperative. Such longitudinal compressive strain could well be caused by plastic squeeze and consequent lateral flow of the adjacent incompetent layers. Simultaneous shear parallel to layering would make the folds tilt and produce the typical drag-fold geometry.

Sheets parallel to y and oriented in the first and third quadrants relative to x and z become folded with axes parallel to y as predicted in the theoretical section. [Note that the shear is performed by sliding the upper wooden block (which cuts the positive z-axis) in the direction of the negative x-axis. In the theoretical consideration of simple shear the movement was in opposite sense.] The axial plane of each fold may or may not be parallel to the long axis of the strain ellipsoid. In rocks this would mean that the axial plane of ptygmatic folds needs not be parallel to schistosity. Unless the sheets are parallel to the direction of principal compressive strain there is a component of shear strain parallel to the sheets; this shear tends to rotate each individual fold. The rotation is particularity evident in layers which make a small angle with the x, y-plane as shown in fig. C, pl. 6. Such ptygmatic folds have the characteristic shape of drag folds.

Wave length or rather arc length and amplitude are related to thickness of the plasticene sheets as shown in figs. B, C, D, pl. 3, and E, F, G, D, pl. 4.

Fig. 12 shows five steps in a sequence of deformation of three sheets originally parallel to y but inclined at various angles to the x-axis. It is seen how shortening as well as intensity of folding depends upon angular attitude in a fashion expected theoretically. However, it is noteworthy that the magnitude of shortening or lengthening is somewhat less than geometrically ideal for some of the sheets. Thus, for example, the stretching of the steepest sheet, which in the final step of strain fractured to two boudins, is considerably less than predicted by the simple shear geometry. Likewise, the shortening of the least steep sheet is smaller than geometrically ideal. The chief reason for this is that the longitudinal strain in the putty adjacent to the competent sheets is smaller in magnitude than it would have been in this site if the relatively competent plasticene sheets were absent. The pattern of striation on the putty surface as well as deformed circular marks close to folded sheets demonstrate the inhibiting effect which the competent sheets have on the compressive strain in the adjacent putty, see figs. B, pl. 5, and B–F pls. 3 and 4.

Sheets parallel to y and inclined against x or z in second and fourth quadrants were always lengthened, often without break but sometimes producing pinch-and-swell or boudinage structures. Figs. A, B, C. pl. 2, and fig. 12 show sheets with this orientation. The two steepest sheets in fig. 12 have been rotated from the compression quadrant to the extension quadrant. It is interesting to note how the sheets firstly buckle into folds, then become stretched out, and ultimately fracture to boudins. Actually, the middle sheet, which is not broken in the figure, developed a fracture close to its middle when strained beyond stage 5. For sake of clarity this state is not shown in the figure.

A few tests with embedded plasticene sheets parallel to the free surface of the putty (parallel to the x, z-plane) were performed. The sheets developed folds essentially parallel to the long axis in the strain ellipsoid.

Fig. D, pl. 6 shows the result of simple shear on two sheets embedded at oblique angle to all three coordinate axes. The putty above the sheets has partly been removed to clear the folded surface of the sheets. Sheets with such oblique orientation do of course not have

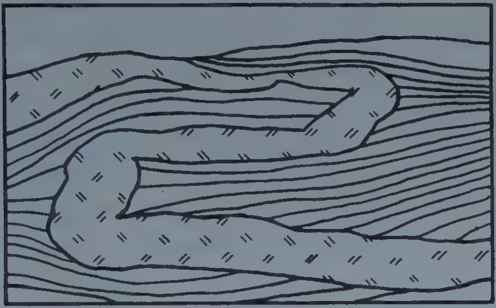

Fig. 13: Ptygmatic vein traced after photo (see Ramberg 1952, p. 257). Note the gneiss structure close to the convex side of the folds.

parallel fold axes in spite of simulteanous deformation in response to the same homogeneous strain in the host rock. The geometry of strain performed in this experiment would correspond to the strain on the flank of a major fold with axis parallel to y ($y = b$). Hence neither of the fold axes of the ptygmatically deformed sheets (veins) would be parallel to the major fold axis in the complex. Structures of this kind in rocks may at first glance erroneously be considered indicative of several periods of deformation.

Details of the nonhomogeneous strain adjacent to the ptygmatic folds are revealed by distortion of original circles impressed on the putty surface across the sheets and close to them. The crucial features of the pattern of this contact strain are the same as those of the contact strain adjacent to many natural ptygmatic veins such as that showned in fig. 13. It is interesting that this kind of contact structure of ptygmatic veins has been considered by G. Wilson (1952, p. 16) as evidence of primary folding caused by forceful injection of a very viscose melt in a less viscose host rock. We see here that such contact structures develop by secondary strain.

It is significant that the nonhomogeneous contact strain is confined to a rather narrow zone along the ptygmatic sheets. That means that the foldings of neighbor veins do not influence each other unless they are less than say one wave length apart. If ptygmatic folds for example were shear folds due to shearing fore and back along the schistosity plane of the host rock there would be a close parallelism between folded veins over a sizable portion of a given schist or gneiss.

Wave length and amplitude should be the same independent of individual vein thickness. This is not what one finds in nature.

Some observations of particular significance for the mechanism of folding as discussed in the following section shall now be mentioned,

If somewhat more than usual care was taken to produce plasticene strips with even thickness throughout their entire length at the same time as the enclosing putty was as homogeneous and free of lumps as possible, it was found that the resulting ptygmatic folds were very uniform in size and shape. Example of this is seen in pls. 3, 4. Most irregularities in the folds as found in the photos are due to heterogeneities in the putty (which often contained remnants of plasticene from previous tests), irregular thickness of the plasticene strips, and/or non-uniform strain due to various boundary effects.

If the ends of the plasticene strips were freely floating in homogeneous putty rather than touching the wooden blocks or some erratic inclusion in the putty, folding was always found to start along a central segment of the strip leaving the end portion seemingly straight and unaffected. In case several waves were produced along the active central segment, it was noted that they appeared simultaneously in the form of equal sized small-amplitude folds whose amplitude gradually increased. The wave length of course decreased correspondingly, but the length of arc along a fold remained constant (see also p. 148). Only in response to increased force applied in the shear motion of the wooden blocks would new folds outside the initial central active segment.

It was noted that the straight terminal segment unaffected by initial folding increased in length with increasing thickness of the strip; see for example fig. D, pl. 4. Of course, if the strip was thick enough it would not fold at all implying that the active central segment has shrunk to zero and the whole strip is occupied by the inactive straight end segments.

The ptygmatic folds could easily be unfolded again by reversing the direction of sliding of the wooden block as demonstrated in fig. G, pl. 4.

In addition to the various types of distortion of embedded competent sheets, other features of consequence for strain structures in rocks were revealed during the experiments with simple shear. A set of quite uniform shear fractures developed in the putty at about

20°–25° angle with the x, y-plane. The acute angle opened in the directions of shear motion, Pls. 1, 2, 4. Now maximum finite shear strain is parallel to the two circular cross-sections of the strain ellipsoid, one being fixed in the x, y-plane, the other rotating from the y, z-plane into the second and fourth quadrants. The experimentally produced fractures deviated considerably from either of these directions. The same holds for some fractures developed in shear experiments by Riedel, 1929. The orientation of the fractures is such that a component of extensive strain occurs perpendicular to the fractures. If deformation is continued after the fractures first appear on the putty surface, one finds that many of them open up a little at the same time as lateral shear displacement is evident. Thus the actual fractures are hybrids between "ideal" shear fractures which have no transversal extensive or compressive strain and "ideal" tension fractures along which shear strain vanishes.

On the surface of the strained putty a very conspicous set of striations developed parallel to the long axis of the strain ellipsoid. This striation really seems to be a minute folding of the thin surface crust of the putty. It is parallel to the direction of slaty cleavage as developed in incompetent rocks sheared between competent beds. The above mentioned hybrid shear-tension fractures were oriented quite opposite to the slaty cleavage in natural layered rocks. The experiments thus support the theory that slaty cleavage develops parallel to maximum finite extensive strain in rocks rather than parallel to any of the shear directions.

Some of the putty batches used contain remnants of plasticene from previous experiments. These remnants were generally nebulous and contorted in irregular manner. During simple shear experiments some of these contorted nebulous remnants developed a kind of planar structure very similar to axial plane cleavage in slate.

Experiments with pure shear and various three dimensional strain.

It was found difficult to maintain homogenous strain under these types of deformation because of friction along the interfaces between yielding putty and the rigid plates which in most cases were used to transmit stresses. The resulting strain pattern of the embedded plasticene sheets were therefore generally somewhat distorted by the nonhomogeneous component of the strain. A typical example of such distortion is seen in fig. D, pl. 5, which are results of simple compression between parallel plates. Because of friction between plates and putty, flow and longitudinal strain parallel to the plates are more intensive in the middle than along the margins. This gives rise to the outward bend in the plasticene sheets. The small folds are however due to the compressive strain. To reduce the effect of frictional drag, the plates were intermittenly loosened from the putty body at short time intervalls during the strain experiments.

In all experiments under this heading the plasticene strips or sheets were entirely enclosed in the putty and their final deformation studied in cuts through the putty cakes.

Results of various strain tests are shown in pls. 2, 5, 6. Explanation is given in the figure texts. Some of these strain features are worthy of special mention inasmuch as they did not develop in the simple shear tests.

Most striking is the result of uniform compression in the x, y-plane and corresponding extensive flow in z. Plasticene sheets oriented parallel to x, y became buckled in a very complex fashion in response to such strain. Fig. B, pl. 6, shows an example of that. Cuts parallel to z through such a buckled plasticene sheet will reveal rather ordinary ptygmatic pattern, but cuts perpendicular to z may show veins with closed outlines much like contour lines on a map of a mountainous terrain, fig. 10.

Under this type of strain two perpendicular veins both parallel to the z-axis will for example develop ptygmatic folds simultaneously

Compression parallel to z and simultaneous uniform extension in the x, y-plane give rise to two-dimensional pinch-and-swell or boudinage structures of competent sheets oriented parallel to the x, y-plane.

Fig. A, pl. 5 shows the result of pure strain on putty with two enclosed plasticene sheets, one originally parallel to the x, y-plane and one originally almost parallel to z. The one sheet buckled into folds at the same time as the other ruptured to boudins. This demonstrates how, in natural rocks, ptygmatic foldings and pinch-and-swell or boudinage structures may well form in response to one and the same period and geometry of strain in a rock complex.

Some of the figures, for example D, pl. 5, shows that wave length of the folds to some extent depends upon thickness of sheets.

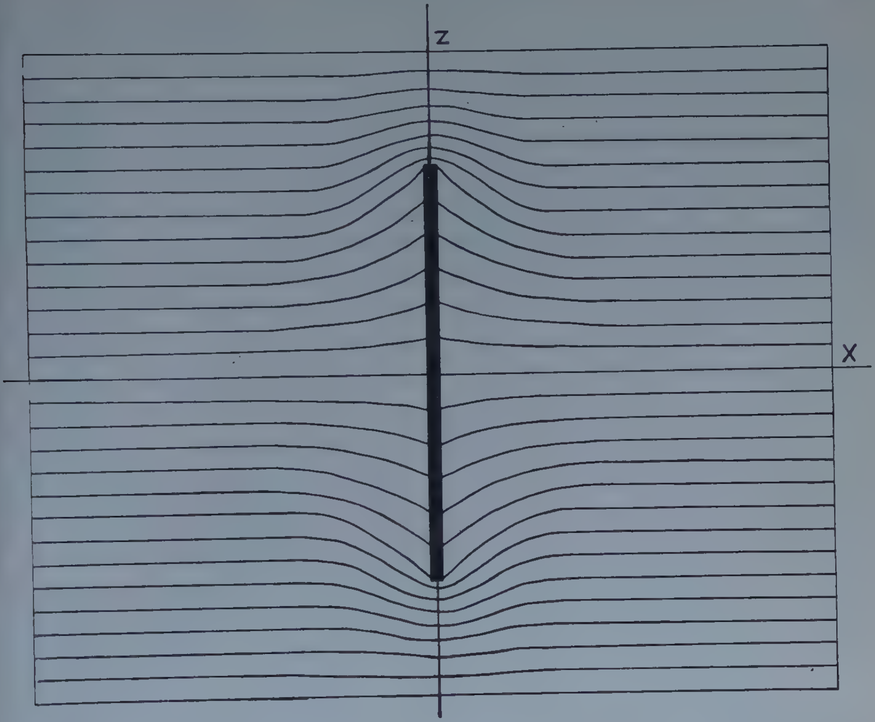

Fig. 14: Deformation of original horizontal marker lines in a viscose or plastic body surrounding a stiff plate oriented parallel to the maximum compressive stress in pure shear. Compressive strain is about 33 %.

Forces and mechanism of ptygmatic folding.

The evolution of ptygmatic folding of veins can in some respect be compared with buckling of metal plates under compressive forces acting on the edges and parallel to the plate. Such buckling phenomena are of practical importance for mechanical engineers and have consequently been studied rather thoroughly within the elastic region of metals and other construction material. There are, however, several conditions which make the evolution of natural and experimental ptygmatic folds quite unlike the simple buckling tests of the engineers. For one thing the natural and experimental ptygmatic folding takes place largely within the plastic domain of the materials. Secondly the ptygmatically folded veins or plasticene strips are surrounded by material which is almost equally "firm" or viscose as the folded bodies themselves. In practical buckling tests the viscosity of the surrounding media is negligible because the test pieces are generally surrounded by air or liquids. Thirdly the buckling forces are usually applied on the edges of the test sheets in the form of compressive pressures parallel to the sheets. This creates a uniform compressive plane stress throughout the entire sheet prior to buckling. In natural and experimental ptygmatic folding, however, stress in the sheet is chiefly caused by drag between the yielding host material and the competent sheet. This drag creates a non uniform plane stress which increases continuously from the edges toward the center or central cross section of the sheet. A brief analysis of this plane stress follows.

Consider a vein which is parallel to maximum compressive stress, σ_z, in a homogeneous plastic rock of great extension in direction perpendicular to the vein, see fig. 14. Let the geometry of the strain be of pure shear type. We shall also assume that the host rock behaves like a very viscose Newtonian substance — i.e. a medium which has no yield point and in which rate of strain is proportional to applied stress. This assumption is reasonable for the slow creep of rocks at stresses below their elastic limit. At some distance away from the vein in x and z directions the host yields by simple compressive strain at constant rate parallel to z. In the host rock outside the disturbing effect of the stiff vein there is no shear strain parallel to z which is the direction of principal compressive strain. As one approaches the competent vein, however, shear strain parallel to z becomes more and more pronounced, and in direct contact with the vein the rate of shear parallel to z reaches maximum values whereas the rate of compressive strain along z approaches zero provided that slip does not occur along the boundary. In a Newtonian substance rate of shear strain is proportional to shear stress. Consequently the surface of the vein is effected by a shear stress parallel to z and directed toward the central cross section (x, y-section) of the veins.

Because of symmetry the shear stresses on either side of the vein are identical in magnitude and sense at given distance z from the center. The relative magnitude of shear strain in the neighborhood of the vein is indicated in fig. 14 which shows how a set of originally even-spaced horizontal parallel marker lines have changed in the course of strain through an arbitrary time interval. The figure shows a case of about 33 % compression parallel to z outside the contact

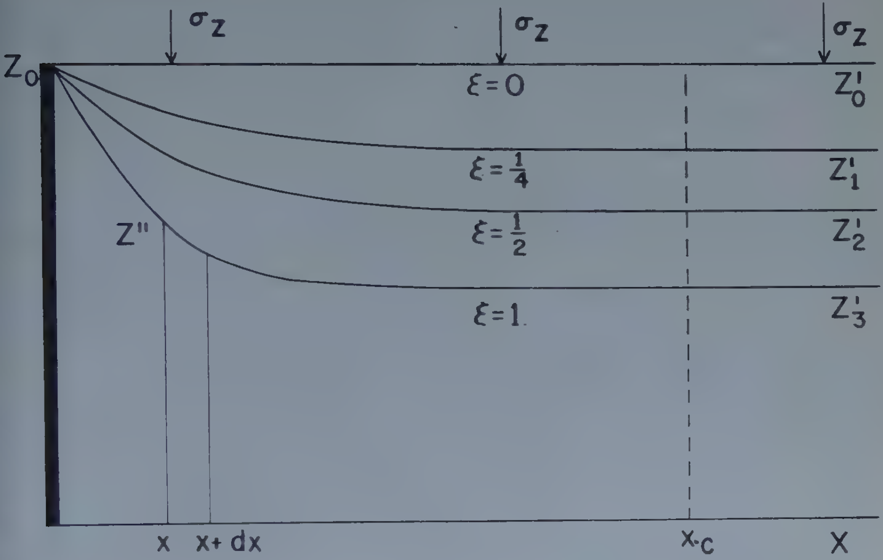

Fig. 15: Deformation of a viscose or plastic body adjacent to a stiff plate. Only half of the plate, which is parallel to the *y*, *z*-plane, is shown (the black body along the *z* axis). Origin of the coordinate system coincides with the center of the plate. Maximum compressive stress is parallel to *z*. The curves $z_0 - z'_1$; $z_0 - z'_2$; $z_0 - z'_3$ represent deformed verson of the line $z_0 - z'_0$ after compressive strain parallel to *z* in the host material outside the contact effect of the plate being $\varepsilon_z = 1/4, 1/2$ and 1. ε_z is defined as $\frac{z_0 - z^1_n}{z^1_n}$.

effect of the competent vein. Vein and host rock (plasticene and putty) are assumed to be completely welded together to prevent slip along the contact. The spacing between the marker lines is therefore unchanged at the vein contact. The drawing is based on actual strain tests with putty and plasticene sheets too strong to buckle, and especial care has been taken to reproduce the correct relative shear angles at the contact. It is found that the tangent of the shear angle $\left(\gamma = \frac{\partial z}{\partial x}\right)$ is closely proportional to the distance *z* from the center, a condition which is reasonable on theoretical grounds as well. Now the finite shear strain at every point along the contact developed in the course of the same time interval; we therefore conclude that the average rate of shear strain at a given contact point was proportional to the finite shear angle at that point. Since rate of shear strain is proportional to shear stress in a Newtonian substance, it follows that shear stress along the vein surface is proportional to the distance from origin:

$$\tau_z = c\, z \tag{37)}$$

where c is a constant depending upon the viscosity of the host material and rate of compressive strain in the host outside the contact zone of the vein.

We shall now develop an approximation equation for shear strain at the ends of a thin, stiff vein (i.e. at $z = z_0$) as a function of compressive strain parallel to the vein in the enclosing rock outside the contact zone. This will enable us to express the shear stress along the vein contact in terms of compressive stress on the host rock. Let the curve $z_0\, z'$ in fig. 15 represent the deformed version of the horizontal line $z = z_0$ after unit time interval, in which the strain rate has been constant. We shall assume that the stress in the z direction is constant $= \sigma_z$* along the curve $z_0\, z'$. Consider now the column $z'' \times y \times dx$ where y is unity. This column has been compressed from z_0 to z'' in unit time. At the same time shearing has taken place along the sides of the column. Along the vertical side at x the shear varies from 0 at $z = 0$ to $\frac{\partial z}{\partial x}$ at $z = z''$. Assuming that shear is a linear function of z the mean shear along this side is $\frac{1}{2}\,\frac{\partial z}{\partial x}$. This shear has developed in unit time and therefore resists the compression of the column by a force $= -\frac{1}{2}\,\mu\, z\,\frac{\partial z}{\partial x}$ where μ is the viscosity of the material. Along the other side of the column at $x + dx$ the shear varies from $\frac{\partial\left(z + \frac{\partial z}{\partial x}\,\partial x\right)}{\partial x}$ at $z = z''$ to zero at $z = 0$. We shall again assume that the mean shear along this side is $\frac{1}{2}\,\frac{\partial\left(z + \frac{\partial z}{\partial x}\,\partial x\right)}{\partial x}$. This shear helps to compress the column by a force $= -\frac{1}{2}\,\mu\, z\,\frac{\partial\left(z + \frac{\partial z}{\partial x}\,\partial x\right)}{\partial x}$.

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* In the following discussion compressive stress and strain are considered positive.

The total compressive force effecting the column is then $\sigma_z \, \partial x - \frac{1}{2} \mu z \frac{\partial \left(z + \frac{\partial z}{\partial x} \partial x\right)}{\partial x}$ which is balanced by the resistence force, $-\frac{1}{2} \mu z \frac{\partial z}{\partial x} + 3 \mu \frac{z_0 - z}{z} dx$ where the last term represents resistance of a viscose body against simple compression. The constant μ equals the viscosity coefficient since the strains have developed at constant rate through unit time interval.

Equating the compressive and the resistive forces gives:

35)
$$\frac{1}{2} z \mu \frac{\partial^2 z}{\partial x^2} = \sigma_z - \frac{z_0 - z}{z} 3 \mu, \text{ or}$$
$$\frac{\partial^2 z}{\partial x^2} = 2 \left(\frac{\sigma_z}{\mu} + 3\right) z^{-1} - 6 z_0 z^{-2}$$

of which the first derivative can be found by integration: (See Forsyth, 1943, p. 88)

36)
$$\frac{\partial z}{\partial x} = \pm \sqrt{4 \left(\frac{\sigma_z}{\mu} + 3\right) \log_e z + 12 z_0 z^{-1} + A}$$

where A is a constant of integration.

According to the model fig. 15 we realize that both $\frac{\partial^2 z}{\partial x^2}$ and $\frac{\partial z}{\partial x}$ equal zero at $x > x_c$, where x_c is the distance beyond which the contact effect of the vein vanishes. In the region $x > x_c$, $z = z'$ where z' is determined by the full compressive strain, $\varepsilon_z = \frac{z_0 - z'}{z'}$, in the host rock. Since $\frac{\partial^2 z}{\partial x^2} = 0$ at $z = z'$, eq. (35) gives:

38)
$$\frac{\sigma_z}{\mu} = 3 \frac{z_0 - z'}{z'} = 3 \varepsilon_z \, *$$

for the region $x > x_c$, or $z = z'$.

The constant A is determining by setting $\partial z / \partial x = 0$ at $z = z'$ in eq. (36):

39)
$$A = -12 (\varepsilon_z + 1) \log_e \frac{z_0}{1 + \varepsilon_z} - 12 (\varepsilon_z + 1)$$

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* Note that according to our definition compressive strain has positive value.

in which the expressions $\frac{\sigma_z}{\mu} = 3\,\varepsilon_z$, and $z' = \frac{z_0}{\varepsilon_z + 1}$ have been introduced. Eq. (36) then becomes:

$$40)\qquad \frac{\partial z}{\partial x} = \pm \sqrt{12\,(\varepsilon + 1)\,\log_e \frac{(\varepsilon + 1)\,z}{z_0} - (\varepsilon + 1)\,12 + 12\,\frac{z_0}{z}}$$

Integration of this equation results in a family of deformation curves running from z_0 at the end of the vein to $z'_1, z'_2, \ldots\ldots z'_n$ at $x > x_c$. Each value of the compressive strain, ε_z, gives one member of this family of curves ending at z'_1, z'_2, etc. Geometrically integrated curves for $\varepsilon_z = 1/4$, $1/2$, and 1 are plotted in fig. 15.

It is interesting to see that the deformation curve from the ends of the vein for 33 % compression (i.e. $\varepsilon_z = 1/2$) in the putty-plasticene experiment fig. 14, coincides almost exactly with the theoretical curve for $\varepsilon_z = 1/2$ (fig. 15). This is particularly true for the region close to the contact of the vein. That means that the shear-strain angle at the vein contact is rather accurately expressed by eq. (40).

It is worth noting that the contact of the competent vein — i.e. the region in which $\partial z/\partial x \lessgtr 0$ — has a limited extension in the x-dimension. x_c appears to equal z_0 rather closely. This is in good agreement with the experimental strain results as shown in fig. 14.

One of the most significant applications of eq. (40) lies in its ability to determine the contact shear strain, $\partial z/\partial x$, at the ends of the vein ($z = \pm\, z_0$, $x = 0$). This is achieved by setting $z = z_0$ at given ε_z-values. Table I* gives correlated values for $\gamma_\delta = \partial z/\partial x$ at $z = z_0$ and $\varepsilon_z = \frac{z_0 - z}{z}$ as determined by eq. (40). It is seen that the shear strain at the ends of the vein is very nearly proportional to the compressive strain parallel to z in the host. This means that the rate of shear strain at the contact of the vein is proportional to the rate of compressive strain in the host at $x > x_c$.

The contact shear stress along the vein parallel to z at the ends can now be expressed in terms of compressive stress, σ_z, in the host by the following relations: For rate of compressive strain in the host rock outside the contact zone: $\sigma_2 = 3\mu\,\dot{\varepsilon}_z$ where $\dot{\varepsilon}_z$ is rate of strain, $\varepsilon_z/\Delta t$. Hence: $\mu = \frac{\sigma_z}{3\,\dot{\varepsilon}_z}$.

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* For Table I, see p. 151.

For shear strain at the contact at z_0:

41) $$\tau_z = \mu \dot{\gamma}_0$$

where $\dot{\gamma}_0$ is rate of shear strain, $\gamma \cdot / \Delta t$, and τ_z is shear stress, hence:

42) $$\tau_z = \frac{\sigma_z}{3} \frac{\gamma_0}{\varepsilon_z}$$

According to table I γ / ε_z is almost constant for all ε_z. The mean value, 2.340, is inserted in eq. (42):

43) $$\tau_z = 0.78\, \sigma_z$$

We have for simplicity assumed σ_x and σ_y to be zero. This is very unlikely in rocks, hence σ_z in equation (43) should be replaced by $\Delta\, \sigma_z$ which is the difference between σ_z and σ_x:

44) $$\tau_{z0} = 0.78\, \Delta\, \sigma_z$$

It follows from eq. (40) that the contact shear-strain rate — and consequently the shear stress — at the ends of the vein is independent of the vein length, $2\, z_0$, at given compressive-strain rate in the host.

The equations above do not show how τ_z varies along the vein contact, they only determine τ_{z0} at the terminals. However, we have concluded from experiments that the shear-strain rate, and consequently the shear stress, along a given sheet or vein is nearly proportional to distance from center. Since the shear stress at z_0 is $\tau_{z0} = 0.78\, \Delta\, \sigma$ the contact shear stress at any point, z, is

45) $$\tau_z = 0.78 \frac{\Delta \sigma_z}{z_0} z$$

where z_0 is the half-length of the vein.

Due to this drag then, the longitudinal compressive force parallel to z across a x, y-section at distance z from center in the vein is as reckoned per unit length in y:

46) $$D = 2 \int_{z_0}^{z} \partial F = -2 \int_{z_0}^{z} \tau_z\, \partial z = -1.56 \frac{\Delta \sigma_z}{z_0} \int_{z_0}^{z} z\, \partial z$$

$$= 0.78 \frac{\Delta \sigma_z}{z_0} (z_0^2 - z^2)$$

The factor 2 comes from the fact that equal stress acts on either side of the vein.

In addition to the drag-induced force there is pressure on the edges at $z = \pm z_0$ which results in an uniform compressive force throughout the entire length of the vein. Per unit length in y-dimension the edge force is:

$$F_e = \Delta\, \sigma_2{}'\, x_0 \tag{47}$$

where $\Delta\, \sigma'_z$ is the difference $\sigma'_z - \sigma'_x$ in the host rock at the end of the vein, and x_0 is its thickness. Incidentally, one notes that σ'_z is somewhat greater than the compressive stress along z in the host outside the contact effect of the vein.

The total longitudinal compressive force across an x, y-section of the vein at z is then per unit length in y:

$$F = F_e + D = \Delta\sigma_z{}'\, x_0 + 0.78 \frac{\Delta\sigma_z}{z_0} (z_0{}^2 - z^2) \tag{48}$$

This is the force which buckles the vein if the force exceeds the resistance against buckling as offered by the strength of the vein and the supporting effect of the enclosing host rock.

At some distance from the end points in a sufficiently long vein the compressive force can be much larger than the compressive stress in the host rock if it flows under the existing stress.

As an example assume $\Delta\, \sigma_z$ in host rock to be say 100 atm. Let a competent vein oriented parallel to maximum pressure be 400 cm long parallel to z and 20 cm thick. At a section 100 cm from the ends, the compressive force parallel to z in the vein per unit length in y is then:

$$F = 100 \times 20 + 0{,}78 \frac{100}{200} (40{,}000 - 10{,}000) \text{ kg} = 13{,}700 \text{ kg.,}$$

or 13,700/20 atm. = 685 atm.

If vein length is say 1000 cm ($z_0 = 500$ cm) the compressive force 200 cm from the ends is:

$$F = 100 \times 20 + 0{,}78 \frac{100}{500} (250{,}000 - 90{,}000) \text{ kg} = 26{,}960 \text{ kg}$$

or 1,348 atm.

These figures show how pressure can accumulate considerably in competent bodies enclosed in incompetent rocks.

It is convenient to consider the resistance against buckling of an embedded vein in terms of two separate forces.

1. The strength against buckling which the vein possesses because of its elastical and rheological characteristics and its dimensions. This buckling strength of the vein at given P, T-conditions, is defined as the longitudinal compressive force needed to initiate buckling if the enclosing material offers no resistance whatever against the deformation — i.e. if the enclosing material was an ideal fluid without strength and viscosity. The effect of an ideal fluid of this type would only be to keep the vein under a given confining pressure thereby partly controlling the mechanical characteristics of the vein material.

2. The resistance against buckling of the vein as offered solely by the strength and viscosity of the enclosing medium. Sidewise motion of folds into the host material is obviously resisted by the viscose or plastic host.

Let us first consider the buckling strengt of the vein. A plate of elastic material effected on opposite edges by a compressive longitudinal force equal to or greater than its buckling strength, is mechanically unstable in the sense that a very minute transversal force is sufficient to initiate buckling which will not pop back upon withdrawal of the small transversal force. If the plate is unable to yield plastically, the buckled state of the compressed plate is stable unless the compressive force is considerably larger than the buckling strength in which case the plate will collapse. If plastic creep occurs in the plate, however, the amplitude of the buckles will increase with time even if the applied stress does not exceed the buckling strength.

Since the theory of elastic buckling of plates is well understood we shall firstly consider that and then see how it may be modified to account for plastic buckling of the type occurring in ptygmatic veins. Fig. 16 shows a plate which is elastically bend by a force P acting parallel to z. In order that static stability shall exist the moment of the buckling force, $P \cdot \Delta x$ must be balanced by the moment of the stresses in the plate. If P is appreciably greater than this critical value the plate will continue to bend until collapse; if P is smaller, the plate will not bend at all.

The plate is compressed on the concave side of the neutral central surface of no strain, and extended on the convex side of that surface. Because of symmetry the tensile stress on the convex side equals numerically the compressive stress on the concave side at given distance from the central surface. Let the longitudinal stress parallel

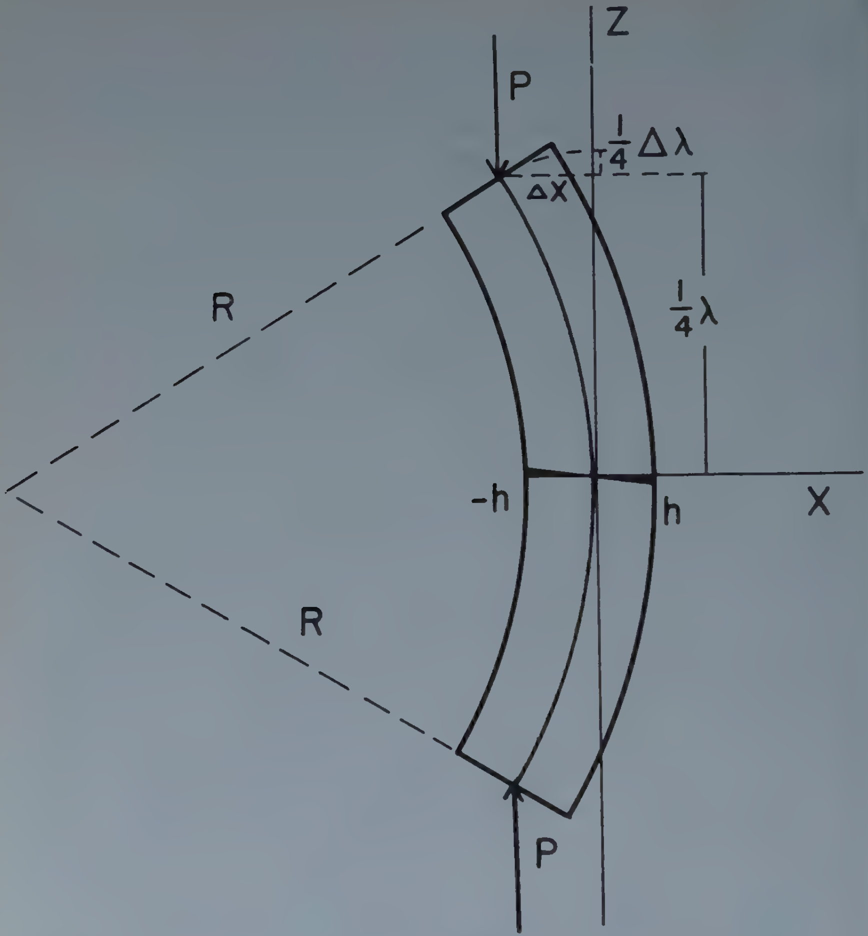

Fig. 16: Cross section of a plate being bent by a force P parallel to z.

to the plate be $\pm\,\sigma$ at distance $\pm\,x$ from the center surface. The moment of the tensile and compressive stresses is then:

$$M = 2\int_0^h \sigma\, x\, d\, x \tag{49)}$$

σ is zero at the central neutral plane and reaches numerically maximum values at $x = \pm\,h$. The change of σ with x in the region $-\,h, +\,h$, is linear:

50) $$\sigma = \frac{\sigma_h}{h}\, x$$

where x varies within the limit $\pm\, h$. σ_h is the stress at the surfaces at $x = \pm\, h$. The value of σ_h is determined by the magnitude of longitudinal strain at $x = \pm\, h$. Provided that the half-wave constitutes a circle arc, this strain is:

51) $$\varepsilon_h = \pm\, \frac{h}{R}$$

where R is the radius of curvature of the buckle. Since the strain is elastic, the ratio between stress and strain equals Young's nodulus, E:*

52) $$\sigma_h = E \cdot \varepsilon_h = E\, \frac{h}{R}$$

For folds with small $\Delta\, x/R$ ratio as we are considering here the following approximate relations can be used:

53) $$\frac{2\, \Delta x}{\frac{1}{4}\, \lambda} = \frac{\frac{1}{4}\, \lambda}{R}, \text{ or } R = \frac{1}{32}\, \frac{\lambda^2}{\Delta x}$$

where λ is the wave length of the folds. Eq. (52) then becomes:

54) $$\sigma_h = E\, h\, \frac{32\, \Delta x}{\lambda^2}$$

The moment due to elastic stress is consequently:

55) $$M = 2 \int_0^h E\, \frac{32\, \Delta x}{\lambda^2} \cdot x^2\, dx = 21.3\, \frac{\Delta x}{\lambda^2}\, h^3\, E$$

Equating the moment of the applied external force, $P \cdot \Delta\, x$, with elastic moment, M, gives:

56) $$P = \frac{21.3\, E h^3}{\lambda^2}$$

P is the force, per unit length in y-dimension, acting in the plane of the plate parallel to z which is required to make the plate of thickness $2\, h$ buckle elastically in folds with wave length λ.

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* For simplicity we shall disregard strain parallel to the fold axis of the buckle. I.e. Poisson's ratio equals zero.

More sophisticated evaluations of buckling strength of plates show that the Poisson's ratio and the ratio between the side dimensions of the plate are of significance as for example discussed by J. J. Stoker, 1941. A comparison of eq. (56) with the more elaborate buckling expressions shows that equation (56) is valid for any number of buckles in plates very elongate in the y-dimension and for square plates if more than 3 buckles are forced to form.

If the plate is not restrained by any outside forces except along the two opposite edges on which pressure is applied, minimum buckling force is attained when the plate makes *one* buckle or *one* half-wave of which the inflection lines coincides with the edges where force is applied. This gives maximum value for wave length in eq. (56) and consequently minimum value for P. If sidewise "popping" is hindered along equal-spaced lines parallel to the compressed edges however, the plate must make several half-waves it if is to buckle at all. As shown by eq. (56) the force necessary to cause multibuckling of this kind increases rapidly with decreasing wave length.

It is interesting to check the applicability of the elastic buckling equation on rocks. For granitic rocks, for example, E is of the order 5.10^{11} dynes/cm^2 ("Handbook", 1942 p. 79–80). $h = 1$ cm, and $\lambda = 20$ cm are reasonable dimensions for an average ptygmatic granitic veinlet. According to these values the buckling force becomes:

$$P = \frac{21.3.5.10^{11}}{400} \simeq \frac{1}{4} 10^{11}$$

dynes per unit length in y-dimension. Since thickness is $2\ h$, this means a compressive longitudinal stress along the vein

$$\text{equal to } \frac{1}{8}\ 10^5 \text{ bars.}$$

This seems an unreasonable high figure, particularly as it must be considered not as absolute stress, but rather as stress difference between the z- and x-and/or y-dimensions.

Let us then attempt to adapt the considerations above to buckling by a slow plastic creep in the plate of the kind likely to take place in veins or other sheet-shaped rock bodies being folded.

The crucial difference between such a mechanism and the elastic case is that in plastic creep rate of strain, $\partial\varepsilon/\partial t = \dot{\varepsilon}$, rather than strain itself is proportional to stress. (For simplicity we assume again that the rocks behave as very viscose Newtonian substances.) It is therefore conceivable that forces much less than the elastic buckling load are capable of causing buckling by a plastic creep mechanism.

Let fig. 16 represent a plastic plate with very small curvature. As in the elastic model the moment due to the external force must be balanced by the stress moment of the slightly buckled plate. Consequently the condition:

$$57)\qquad P\,\Delta\,x = M = 2\int_0^h \sigma\,x\,dx,\ \text{or}\ 2\int_0^h \frac{\sigma_h}{h}\,x^2\,dx = \frac{2}{3}\,\sigma_h\,h^2$$

is still valid. However, rather than causing a static elastic strain the longitudinal stress now gives rise to a slow plastic creep on either side of the neutral middle surface of the fold. The relationship between compressive, respective tensile, stress and creep in a Newtonian substance is:*

$$58)\qquad \sigma_h = 3\,\mu\,\frac{\partial\varepsilon_h}{\partial t}$$

where μ is viscosity for plastic creep and $\frac{\partial\varepsilon_h}{\partial t}$ is rate of compressive or extension strain in the surface layers of the folded plate. According to eqs. (51) and (53):

$$59)\qquad \varepsilon_h = \frac{32\,h\,\Delta x}{\lambda^2}\ \text{(approximately)}$$

For our purpose it is most convenient to relate strain in the surface layers of the plate, ε_h, to relative shortening along the z-dimension of a buckle, ε_z, as caused by the bending. We shall only concern ourselves in the following with the initiation of small-amplitude buckles in a plate. For such initial waves with very mall $\Delta x/\lambda$ ratio, it can be shown geometrically, (fig. 16) that the decrement in wave length, $\Delta\lambda$, as caused by bending, is related to amplitude and wave length as approximately follows:

$$60)\qquad \frac{\frac{1}{4}\,\Delta\lambda}{\Delta x} = \frac{\Delta x}{\frac{1}{4}\,\lambda}.$$

Relative shortening along z for initial waves is defined as

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* This relation holds for "pure shear".

61) $$\varepsilon_z = \frac{\text{length of arc} - \lambda}{\lambda} = \frac{\Delta\lambda}{\lambda}\,, \text{*}$$

hence:

62) $$\Delta x = \frac{1}{4}\,\lambda\,\varepsilon_z^{\frac{1}{2}},$$

which inserted in eq. (59) gives:

63) $$\varepsilon_h = 8\,\frac{h\,\varepsilon_z^{\frac{1}{2}}}{\lambda}\,.$$

Partial differentiation with respect to ε_z and λ results in:

64) $$d\,\varepsilon_h = \frac{4\,h}{\lambda}\left(\varepsilon_z^{-\frac{1}{2}}\,\partial\,\varepsilon_z - 2\,\varepsilon_z^{\frac{1}{2}}\,\frac{\partial\lambda}{\lambda}\right)$$

where $\frac{\partial\lambda}{\lambda}$ can be replaced by $\partial\,\varepsilon_z$.

Accordingly:

65) $$\sigma_h = \frac{12\,\mu\,h}{\lambda}\left(\varepsilon_z^{-\frac{1}{2}} - 2\,\varepsilon_z^{\frac{1}{2}}\right)\frac{\partial\,\varepsilon_z}{\partial t}$$

and:

66) $$P\,\Delta x = \frac{8\,\mu\,h^3}{\lambda}\left(\varepsilon_z^{-\frac{1}{2}} - 2\,\varepsilon_z^{\frac{1}{2}}\right)\frac{\partial\,\varepsilon_z}{\partial t}$$

Replacing Δx by $\frac{1}{4}\,\lambda\,\varepsilon_z^{\frac{1}{2}}$ gives the longitudinal buckling force:

71) $$P = \frac{32\,\mu\,h^3}{\lambda^2}\left(\varepsilon_z^{-1} - 2\right)\frac{\partial\,\varepsilon_z}{\partial t}\,,$$

72) $$P = \frac{32\,\mu\,h^3}{\lambda^3}\left(\frac{\lambda}{\Delta\lambda} - 2\right)\frac{\partial\lambda}{\partial t}\,.$$

P is to be interpreted now as the force parallel to z, reckoned per unit length in y, which gives a rate of buckling or diminishing wave length equal to $\frac{\partial\,\varepsilon_z}{\partial t} = \left(\frac{\partial\,\lambda}{\lambda\,\partial t}\right)$. The equation is only valid for initial waves with very small curvatures; that is to say the quantity, ε_z which represents the relative shortening of the wave length from the

* Compressive strain according to this definition assumes positive values.

plane stage to the slightly buckled stage, is small, say less than $|0.01|$. As one should except, the force increases with rate of buckling. It is also in accord with expectations that P decreases with increasing magnitude of buckling (as expressed by ε^{-1}) because of the increased moment of the force.

It is worth noting that there is no lower limit to the buckling force for plastic buckling within the creep domain below the yield point. If $\frac{\partial \varepsilon_z}{\partial t}$ approaches zero, P also approaches zero provided the $\varepsilon_z \lesseqgtr 0$. The condition $\varepsilon_z \lesseqgtr 0$ means that an ideal homogeneous plate cannot start buckling if it is mathematically straight and the longitudinal compressive forces are symmetric with respect to the central plane. A minute initial bend is necessary to give moment to the longitudinal force. The requirement $\varepsilon_z \lesseqgtr 0$ is an expression of this condition for ε_z, which equals the difference between arc length and wave length divided by wave length, vanishes only for mathematically straight plates. Of course, under natural and experimental conditions the materials are not flawless either in structure or dimentions. There will always be smaller or larger initial curvatures along veins, points of weakness, and/or unsymmetric application of forces to give bending moment to the existing stresses.

We concluded that the forces necessary to buckle granitic veins elastically are much higher than reasonable for crustal stresses. Buckling by plastic creep, however, gives a more likely picture. Let h be 1 cm and λ be 20 cm as in the elastic example. μ having any values between say 10^{14} and 10^{21} poises is possible for solid rocks. As an example a value of 10^{18} poises will be used. Assume then that the original curvature of the vein corresponds to $\varepsilon_z = -0.01$, or that $\frac{\Delta x}{\lambda} = 0.025$ (See eq. (62) p. 144). For P a rather small value of say 100 kg or 10^8 dynes will be chosen. Inserting these values in eq. (71) gives rate of buckling of the fold as expressed in terms of $\partial\varepsilon_z/\partial t$ or $\frac{\partial \lambda}{\partial t\, \lambda}$:

$$10^8 = \frac{32 \cdot 10^{18} \cdot 1}{400} (100 - 2) \frac{\partial \varepsilon_z}{\partial t},$$

or

$$\frac{\partial \varepsilon_z}{\partial t} = \frac{\partial \lambda}{\lambda\, \partial t} \sim 10^{-11}\ \mathrm{sec}^{-1} \sim 0.03$$

per hundred years. In other words, the wave length of 20 cm becomes shortened by some 3 per cent of its original length or 0.6 cm in the course of hundred years. This rough calculation shows that the equation for plastic buckling below the yield point gives reasonable results contrary to the elastic buckling model.

We shall now consider the resistance against initial folding of the vein as offered by the enclosing material because of its firmness and viscosity.

It has unfortunately not been possible to find rigorous treatment of a parallel problem in literature, but a simple analysis shows that the resistance against initial buckling of a thin sheet offered by an enclosing viscose or plastic medium increases with increasing wave length of the folded sheet. As a matter of fact the enclosing rock tends to force a stressed vein to develope an infinite number of fininitisimal folds whereas the strength of the vein itself tends to produce one single half-wave covering the entire vein length. The actual arc length of ptygmatic folds are results of a compromise between these two opposite tendencies.

In analogy with the equation for slow transversal motion of long cylinders or elongate plates in a highly viscose medium, as for example studied by Lamb (1932, p. 616), we shall assume that a relationship of the following kind is applicable to initiation and growth of folds of a plate enclosed in a viscose or plastic substance with viscosity μ:

67)
$$F_x = C\,\mu\,\frac{\partial\,(\Delta x)}{\partial t}$$

where F_x is the force in x-direction, reckoned per unit length along the fold axis, that is needed to increase the amplitude of one half-wave at a rate $\frac{\partial\,(\Delta x)}{\partial t}$. F_x is thus the resistance force connected with the deformation of the host material. Now the buckling force acting on veins as well as their buckling strength have been considered in terms of components in the z directions along the veins. To make F_x comparable with these forces, then, it must be transformed into an equivalent resistance force, F_z, acting in the z-direction along the vein. The transformation is readily done by the law of equal moments:

68)
$$F_x\,\Delta x = M,$$

where M is the moment of the resistance force in x-direction acting on $\frac{1}{4}$ wave.

Assuming that F_x is uniformly distributed over the whole half-wave, the average force per unit area of the wave is $\frac{F_x}{\frac{1}{2}\lambda}$, per unit length along y, and the moment is:

69) $$M = \int_0^{\frac{1}{4}\lambda} \frac{F_x}{\frac{1}{2}\lambda} z\,dz = \int_0^{\frac{1}{4}\lambda} 2\,C\,\frac{\mu}{\lambda}\,\frac{\partial(\Delta x)}{\partial t}\,z\,dz = \frac{1}{16}\,C\,\mu\,\frac{\partial(\Delta x)}{\partial t}\,\lambda.$$

Consequently:

70) $$F_z = \frac{1}{16}\,C\,\mu\,\frac{\partial(\Delta x)}{\partial t}\,\frac{\lambda}{\Delta x}$$

Replacing $\partial(\Delta x)$ and Δx according to eq. (62) and its derivative gives F_z as a function of relative shortening of the wave length, ε_z, and rate of shortening, $\partial\varepsilon_z/\partial t$:

73) $$F_z = \frac{1}{32}\,C\,\mu\,\lambda\,(\varepsilon_z^{-1} + 2)\,\frac{\partial\varepsilon_z}{\partial t},$$

or:

74) $$F_z = \frac{1}{32}\,C\,\mu\left(\frac{\lambda}{\Delta\lambda} + 2\right)\frac{\partial\lambda}{\partial t}.$$

The most significant information contained in this equation is that the resistance force against buckling increases with increasing wave length and increasing viscosity of the host material for given relative shortening and given rate of relative shortening. The effect of the enclosing material is consequently to make the wave length as small as possible. If the vein had no strength itself, the host rock would make the wave length infinitly small. This means that the vein would only be compressed along z and thickened parallel to x, but not folded. That is what happens in the experiments when vein and host have the same viscosity and strength.

The total longitudinal compressive force, as reckoned per unit axial length, necessary to make the vein buckle at a rate $\partial\varepsilon_z/\partial t$, or $\frac{\partial\lambda}{\lambda\,\partial t}$ is then:

75) $$F = F_z + P = \frac{1}{32} C \mu_1 \lambda (\varepsilon_z^{-1} + 2) \frac{\partial \varepsilon_z}{\partial t} + 32 \frac{\mu_2 h^3}{\lambda^2} (\varepsilon_z^{-1} - 2) \, \partial \varepsilon_z / \partial t$$

where the first term represents the resistance offered by the enclosing rock due to its viscosity, and the second term represents the buckling resistance of the vein itself.

Eq. (75) is only valid for folds with small $\Delta x/\lambda$ ratio. In other words the quantity ε_z must be small, say less than 0.01.

If everything except wave length and force is kept constant in eq. (75) it appears that F has a minimum value for a certain wave length. This is the very wave length which will develop when buckling starts. Of course, as buckling continues in response to compressive strain in the host rock, the wave length shortens. The length of arc of the waves, however, remains constant and equal to the initial wave length during the subsequent shortening (provided that the length of the vein does not change during the deformation). The magnitude of the initial wave length can therefore always be determined in the field even on mature ptygmatic folds.

An expression for the initial wave length, λ_i, is readily determined by differentiating eq. (75) with respect to λ and equate to zero:

76)
$$d F = \frac{1}{32} C \mu_1 (\varepsilon_z^{-1} + 2) \, \partial \varepsilon_z / \partial t \, d \lambda - 64 \mu_2 h^3 (\varepsilon_z^{-1} - 2) \, \partial \varepsilon_z / \partial t \frac{\partial \lambda}{\lambda_i^3} = 0$$

Hence:

77) $$\lambda_i = h \sqrt[3]{\frac{2048}{C} \frac{\mu_2}{\mu_2} \frac{(\varepsilon_z^{-1} - 2)}{(\varepsilon_z^{-1} + 2)}},$$

or

78) $$\lambda_i = h \sqrt[3]{\frac{2048}{C} \frac{\mu_2}{\mu_1}}$$

because ε_z is necessarily very small when buckling starts. This expression for initial wave length is in harmony with general field experience. It is generally true that λ_i increases with thickness of the vein and with increasing ratio of viscosity (strength) of vein to viscosity (strength) of host rock.

Introducing λ_i in eq. (75) gives the minimum force necessary to buckle the vein:

79) $$F_{min} \cong \left[\frac{1}{32} C \mu_1 \sqrt[3]{n} + 32 \frac{\mu_2}{\sqrt[3]{n}}\right] \frac{h}{\varepsilon_z} \frac{\partial \varepsilon_z}{\partial t}$$

where

$$n = \frac{2048}{C} \frac{\mu_2}{\mu_1}, \text{ and } \varepsilon_z \gtreqless 0.01.$$

It is interesting to note that the initial wave length, λ_i, does not depend upon the rate of buckling, $\partial \varepsilon_z / \partial t$, or the magnitude of relative shortening, ε_z, at which buckling starts. However, the minimum force at which the initial develop is indeed strongly influenced by both rate of compression and absolute magnitude of compression. This condition is significant for the evolution of ptygmatic folds in rocks because if initial wave length — which is identical to the length of arc of the mature folds — was significantly dependent upon rate of compressive strain in rocks it should probably result in very erratic correlation between thickness and arc length of natural ptygmatic veins. The fact that λ_i is practically independent of relative shortening (if it is small) along z, as expressed by ε_z, at the moment buckling starts indicates that initial wave length is not effected by original small-amplitude bends along an uniform vein. The term ε_z in the equations is an expression of the curvature of such a small bend; see eq. (61) p. 144.

It is unfortunate that the factor C in eq. (67) is unknown because otherwise measurements of length of arc and thickness on ptygmatic folds — which according to the model above are initiated at the minimum compressive force F_{min} — could be introduced in the expression for λ_i (eq. (78)) and thus enable us to determine relative creep viscosities for vein and host rock. It is hoped that future studies will give a more accurate equation for the buckling resistance due to host-rock deformation adjacent to folds.

We are now in position to follow in some detail the evolution of a ptygmatic vein which is oriented parallel to maximum compressive strain in an incompetent rock undergoing pure-shear deformation. Drag along the contact induces a longitudinal compressive force which increases from the ends to the central section of the vein according to eq. (48). This force tends to make the vein buckle. The rate of buckling must be very slow along segments of the vein near its ends where the force is small. However, at a certain distance,

$\varDelta z = z_0 - z_c$ from the ends the accumulated compressive force reaches a certain critical value. This critical value of the force is characterized by being exactly sufficient to make the vein buckle at a rate $\frac{\partial \lambda}{\lambda \partial t}$, or $\frac{\partial \varepsilon_z}{\partial t}$ equal to the rate of compression in the host rock *outside* the contact zone of the vein. The result of this is then that folds with wave length, λ_i determined by eq. (78) are initiated rather suddenly and simultaneously over the central vein segment from $-z_c$ to $+z_c$. During further evolution the folds in the central segment continue to shorten in pace with the host-rock compression outside the contact zone. In other words $\frac{\partial \varepsilon_z}{\partial t}$ or $\frac{\partial \lambda}{\lambda \partial t} = \frac{\partial \varepsilon_{zc}}{\partial t}$ where ε_{zc} is compressive strain in host rock ø *outside* the contact zone. Shortening of folds in the central segment continues in pace with the distant host-rock compression, $\frac{\partial \varepsilon_{zc}}{\partial t}$, for some time because the buckling resistance decreases with increasing amplitude of the folds. As the folds become thightened, however, the resistance against buckling increases mainly because of pinching of host-rock material in the concave portions of the folds. Hence it seems likely that further shortening of the folds in the central segment goes on more slowly than the distant host-rock shortening.

Folding of the end segments of the vein also occurs but at a slower rate because of less compressive force. (Note the straight undisturbed end segments in many of the experimental ptygmatic structures. See pls. 1 to 6).

If the vein is evenly thick and both vein and host rock are homogeneous the initial wave length should not be influenced by force of compression or rate of deformation, as shown by eq. (78). and uniform wave length or rather arc length along the entire vein should be the result.

TABLE I

Relations between $\varepsilon_z = \frac{z_0 - z'}{z'}$ and $\gamma_0 = (\partial z/\partial x)_{z_0}$ according to expression.:

$$\left(\frac{\partial z}{\partial x}\right)_{z_0} = \pm \sqrt{12(\varepsilon+1)\log_e(\varepsilon+1) - 12(\varepsilon+1) + 12}$$

z'/z_0	ε_z	$\delta z/\delta x = \gamma_0$	γ_0/ε_z
0.5	1	2.153	2.153
0.6	0.6667	1.487	2.230
0.666	0.50	1.138	2.276
0.7	0.429	0.9867	2.300
0.8	0.250	0.589	2.356
0.9	0.111	0.2648	2.386
0.95	0.0526	0.1286	2.445
0.98	0.0204	0.0504	2.471
0.99	0.0101	0.02486	2.461
1.0	0.0000	0.000	—
Average			2.340

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PLATE TEXT

Pl. 1: *Simple shear deformation of putty body with embedded plasticene strips.*

A, B, C, and D, represent consecutive steps of a deformation at which the upper wooden block has been moved toward the left. The folded plasticene strip to the right lies in the compression direction. Note that folding starts in the middle of the strip where longitudinal compressive strain is highest. The strip to the left is parallel to one of the two principal shear planes and has neither been folded nor stretched. Hybrid shear-tension fractures are seen in figs. C and D.

Pl. 2: *Deformation of putty bodies with embedded plasticene strips.*

A, B, and C represent consecutive steps on simple shear deformation at which the upper wooden block has been slid toward the left. The plasticene strip to the left is effected by a component of extensive strain and has consequently been stretched with break in the middle where tensile stress is at a maximum. Note the unfolded end segments of the compressed strip to the right. Hybrid shear-tension fractures are visible in fig. C. Fig. D is the result of pure shear deformation (= vertical compression) of a sheet of plasticene embedded in putty. Several cuts of a single compressed cake are shown.

Pl. 3 and 4: *Deformation of putty body with embedded plasticene strips.*

A, B, C, D, E, and F represent consecutive steps on a continuous deformation at which the upper block has been slid toward the left. G shows reversal of the movement after step F. Both strips are parallel to a component of compressive strain, the left strip, however, being a little thicker than the right one. (This difference in thickness may not show up on the photos because the surface of the strips were somewhat smeared during smoothening of the putty prior to deformation). Note the difference in wave length of the two strips, and that folding starts in the central segment of the strips. Fig. H pl. 4 shows different wave length of three buckled plasticene strips with unlike thickness. The thickness of the strips increases from right to left.

Pl. 5: *Deformation of plasticene strips embedded in putty.*

A shows cuts of a cake effected by compression in vertical direction. The black pinching-and-swelling body in the middle was a uniformly thick plasticene layer oriented perpendicular to maximum compression. The gray folded inclined "vein" in the center was a straight plasticene sheet oriented somewhat inclined to compression axis, but yet having a component of compressive strain parallel to itself.

B shows deformation of a plasticene strip and originally parallel vertical marker lines during simple shear.

In *C* a folded plasticene strip and striations parallel to the long axis of the strain ellipsoid are shown.

D shows cuts of a vertically compressed putty cake with two vertical plasticene sheets. Note the difference in wave length as related to different thickness.

Pl. 6: *Deformations of putty with embedded plasticene sheets.*

A: Cuts of vertically compressed putty cake with ptygmatic plasticene sheet.

B: Result of uniform compression in the plane of the picture and elongation perpendicular to the picture. It shows the folded surface of an originally plane plasticene sheet which was embedded in putty and oriented in the plane of compression. The putty was removed from one side of the crumpled plasticene sheet after deformation.

C: Plasticene strip in putty deformed in simple shear.

D: Two plasticene sheets originally completely embedded in putty deformed by simple shear at which the upper wooden block was slid toward the left. The two sheets plunge down at about 30° toward the middle of the photographs as indicated by the arrows on the fold axes. The putty which covered the folded sheets has been removed after the deformation run was completed.

PLATE 1

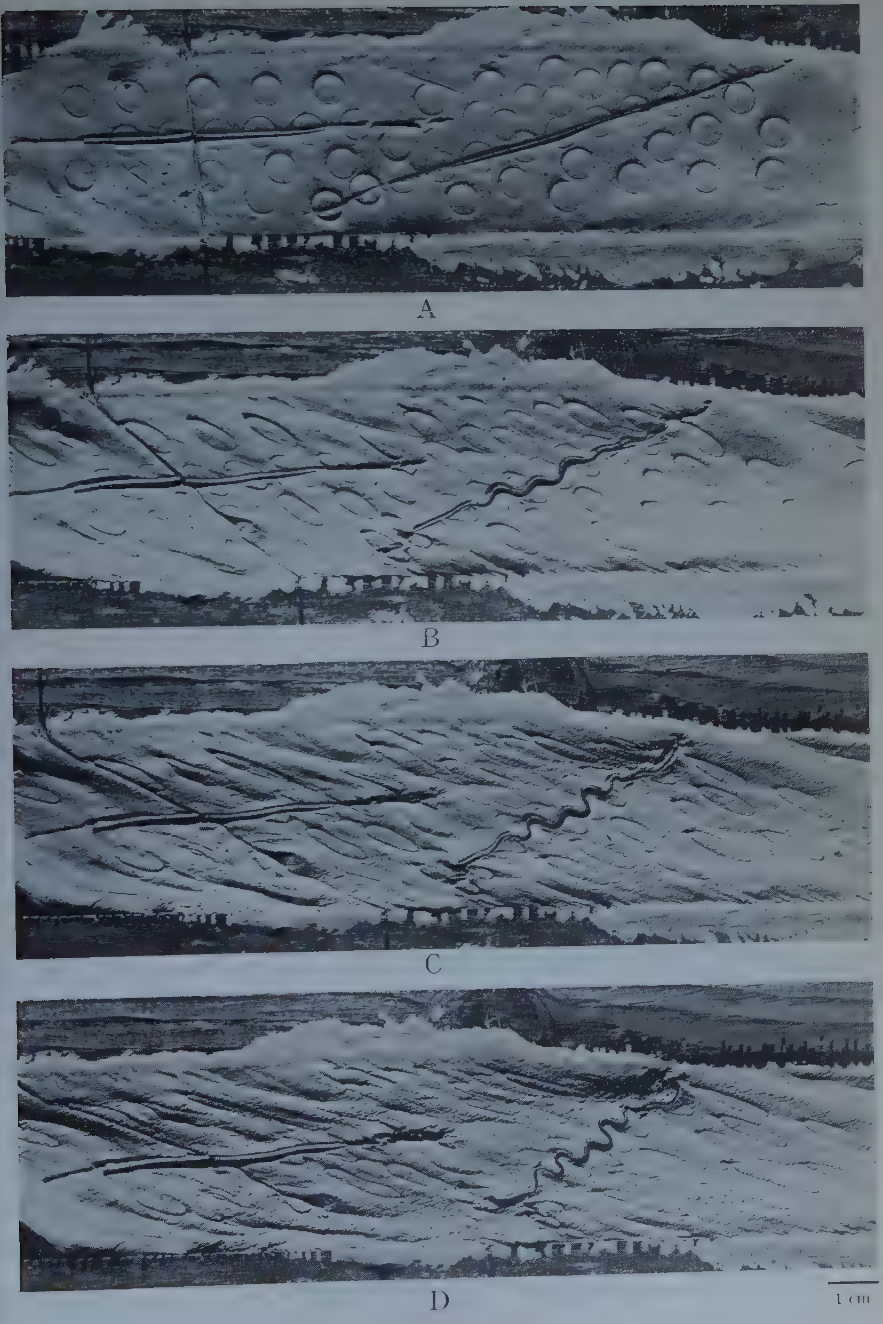
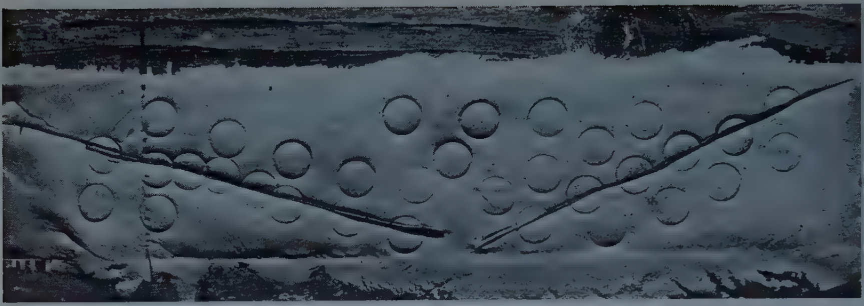
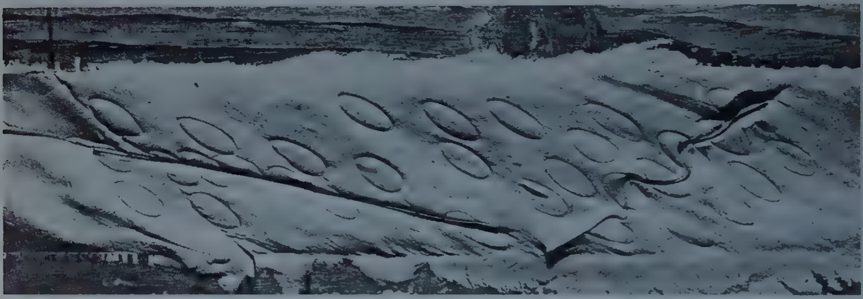


PLATE 2



A



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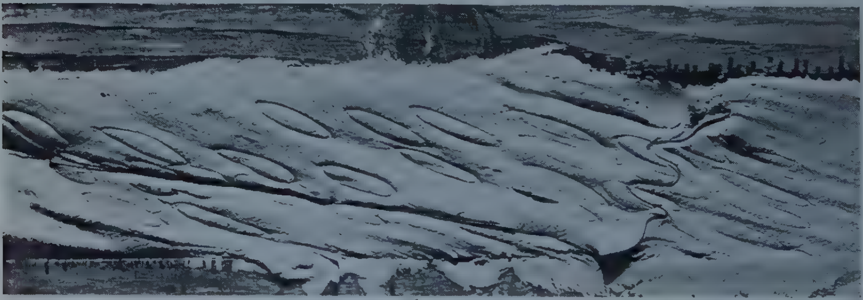

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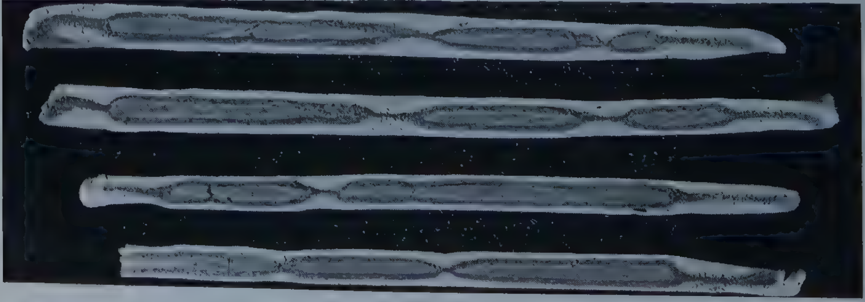

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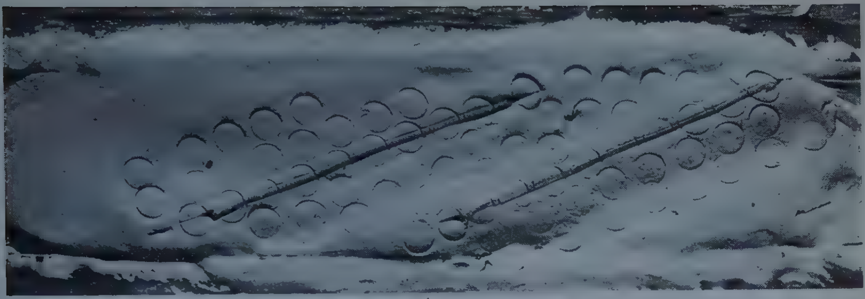

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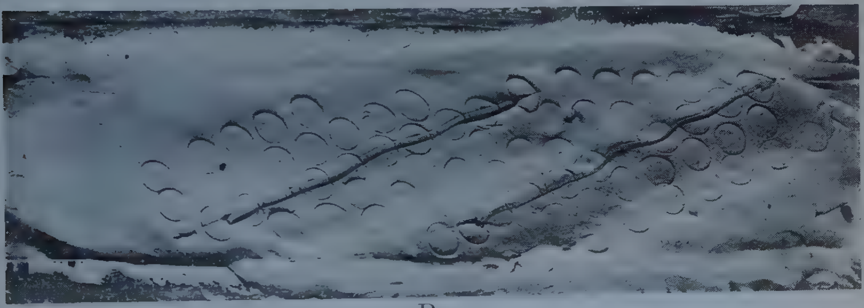

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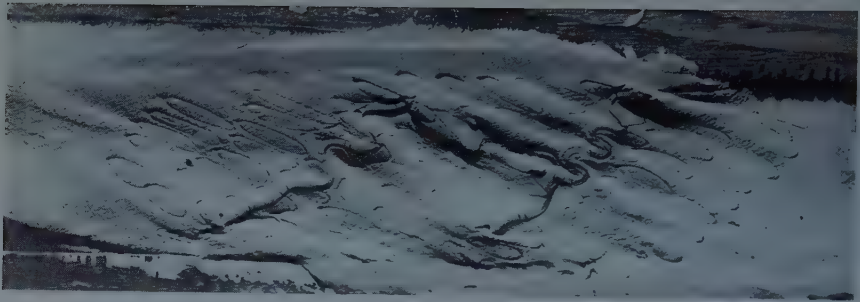

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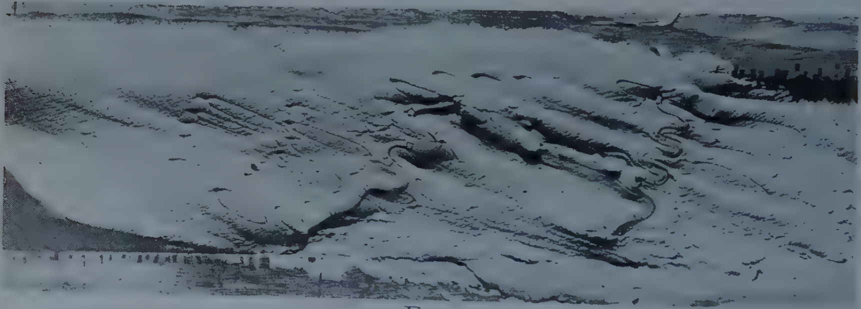

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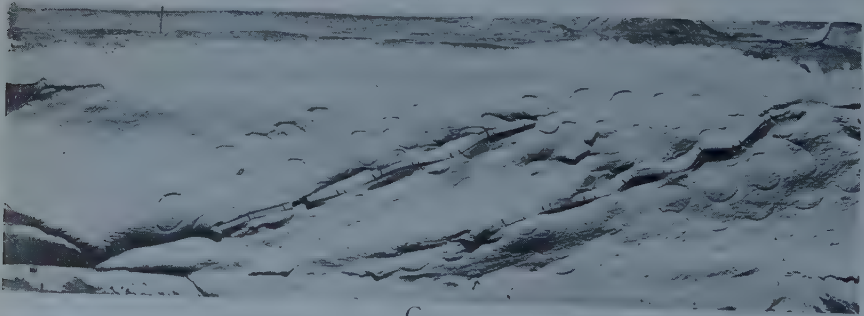

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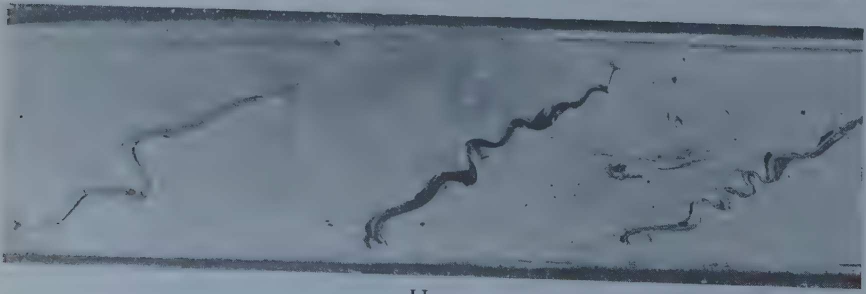

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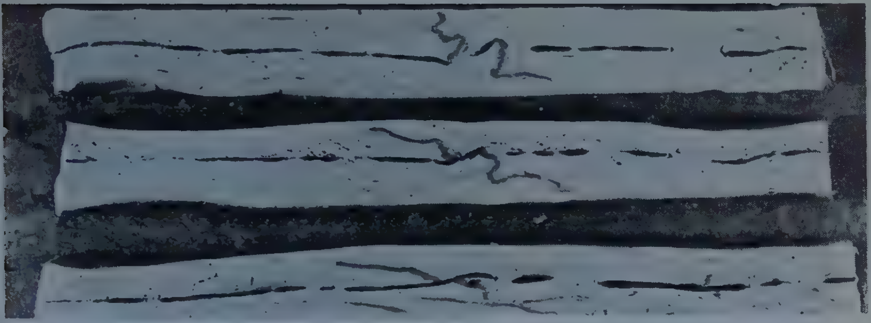

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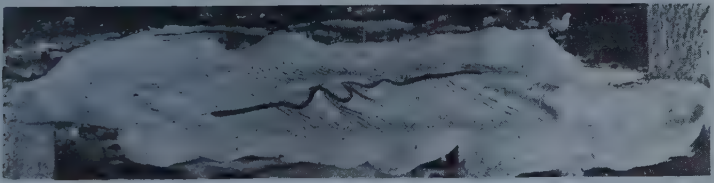

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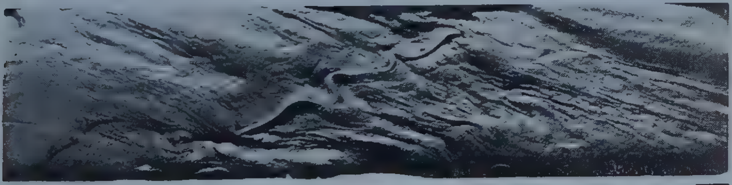

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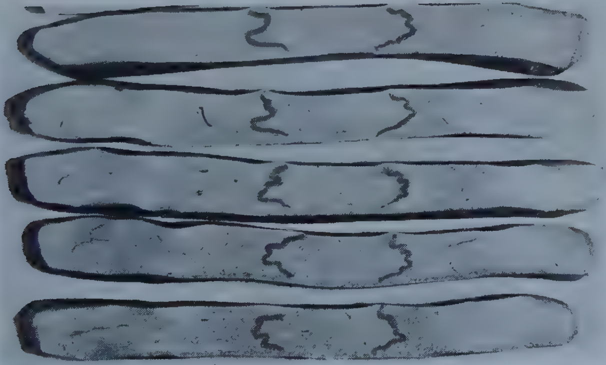

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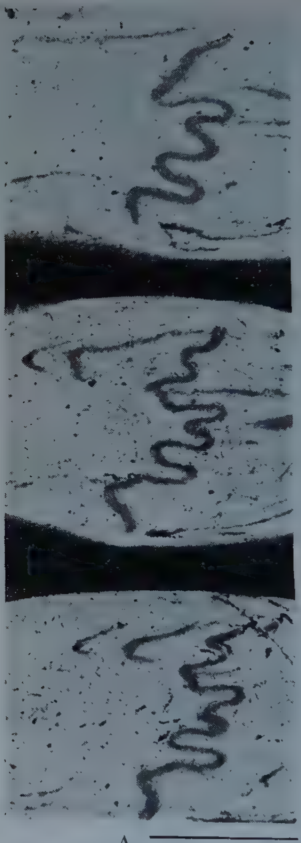
PLATE 6



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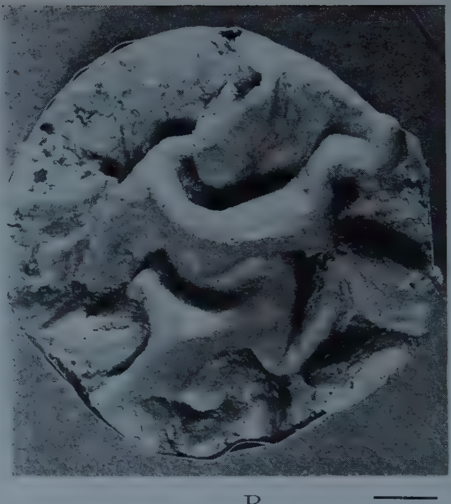

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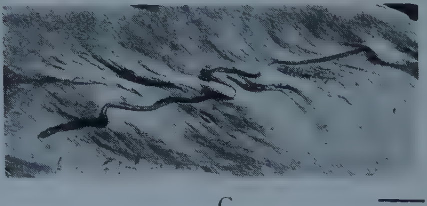

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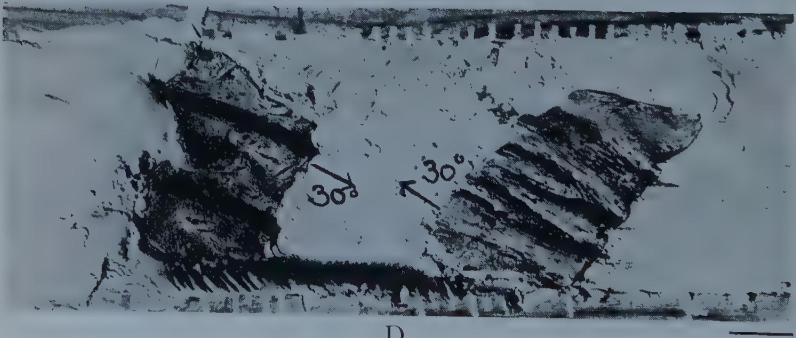

D

RARE TREMADOCIAN TRILOBITES FROM NORWAY

BY

GUNNAR HENNINGSMOEN

(Paleontologisk Museum, Oslo).

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Abstract: Six rare trilobites from the Tremadocian Ceratopyge Limestone in Norway are described. The genus *Pagometopus* (close to *Orometopus*) and the species *Pagometopus gibbus* and *Peltocare modestum* are new. The terminology for the exoskeletal plates of the cephalon is revised. A chart shows the stratigraphic occurrence of the Tremadocian trilobites recorded from Norway.

Introduction and Acknowledgements

During the late autumn of 1958 I collected some trilobites from the Ceratopyge Limestone (substage 3aγ) of Late Tremadocian age at localities north of Slemmestad, 20 km southwest of Oslo. Some of the trilobites proved to belong to rather rare species, two of which had not been described earlier. At several of the collecting trips I was ably assisted by Mr. Frank Nikolaisen, who has presented some of the specimens dealt with in this paper to the Palaeontological Museum in Oslo. There was additional material of some of these species preserved in this museum, and collected after the time of Prof. W. C. Brøgger, who described and made famous the *Ceratopyge* or *Euloma-Niobe* fauna of the Ceratopyge Limestone (Brøgger, 1882; 1896). So far, only 1—12 specimens have been collected in Norway of the species described below, although certain layers of the Ceratopyge Limestone are extremely rich in trilobites and have yielded large collections. Three of the species described here are known also from Sweden, where they likewise are rare. I have preferred to describe the species in the present paper rather than to postpone their description to an eventual general revision of the fauna of the Ceratopyge Limestone in Norway.

The very similar fauna of the Ceratopyge Limestone in Sweden has been described *i.a.* by Moberg & Segerberg (1906), and more recently by Tjernvik (1956) in an important paper, where the generic reference has been changed in many of the species occurring in Norway.

I wish to express my sincere thanks to Miss B. Mauritz for taking the photographs, Miss I. Lowzow for drawing the text figures, and Dr. R. D. Morton for reading the manuscript.

Palaeontological Museum of the University in Oslo, February 1959.

G. Henningsmoen

Terminology

The exoskeleton of the head, or cephalon, consists of 1[1]—6 free exoskeletal plates (*i.e.* plates separated by sutures), like the rostral plate and the cranidium. The cephalon is also divided into a number of other morphological features (areas, furrows, etc.), and for these

---

[1] If the agnostids had no hypostome, otherwise 2—6.

I follow on the whole the terminology applied in the Treatise on Invertebrate Paleontology, except that I use the term *glabella* in its original sense, *i.e.* to include the occipital ring, and the term *dorsal furrow* instead of axial furrow. Furthermore the symbols *S1*, *S2*, etc., and *L1*, *L2*, etc. are used for the glabellar furrows and lobes, respectively, counted from the rear (cf. Jaanusson, 1956; Henningsmoen, 1957). I shall, however, propose a few new terms for exoskeletal plates of the head (text fig. 1).

In opisthoparian, gonatoparian, and proparian trilobites (*i.e.* forms with functional facial sutures) the cephalon consists of the cranidium, librigenae (free cheeks), hypostome[2], and in many forms also the rostral plate (rostrum), and, rarely, the metastome.

In hypoparian trilobites the cephalon has a dorsal plate, a ventral plate called the doublure or inner lamella, and a hypostome. The dorsal plate is often referred to as the cephalon. To avoid this ambiguity of the term cephalon, I propose to call the dorsal plate (glabella + gena) the *genicranium* (pl. *genicrania*); from Latin *gena* (cheek) and *cranium* (skull). The ventral plate I propose to call the *doublural plate*, which I believe is better than using the more general term doublure, especially in trilobites where this ventral plate carries the genal spines. Thus the doublural plate of *e.g. Tretaspis* consists of the doublure *and* the genal spines. Such forms might be regarded as opisthoparians, the librigenae being united and consisting of the doublure and genal spines only. Even so, the terms genicranium and doublural plates seem better for these forms than cranidium and united librigenae. There is an even morphological transition from trilobites with facial sutures to hypoparian ones with marginal suture, and thus also from a cranidium to a genicranium. I suggest to restrict the use of the term cranidium to trilobites where the sutures cut across the genal fields, *i.e.* inside the border furrow. I would then use the term genicranium also in those conocoryphids where the suture cuts off the genal spines and a slice of the border on each side. At least in some conocoryphids the doublural plate is separated into two doublural plates by the intervening rostral plate.

In olenellid trilobites the cephalon consists of the rostral plate,

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[1] There is some inconstancy as to whether the hypostome (and metastome) should be regarded as a part of the cephalon. I see no good reason for not doing so, more so as the hypostome in some trilobites is fused with the rostral plate.

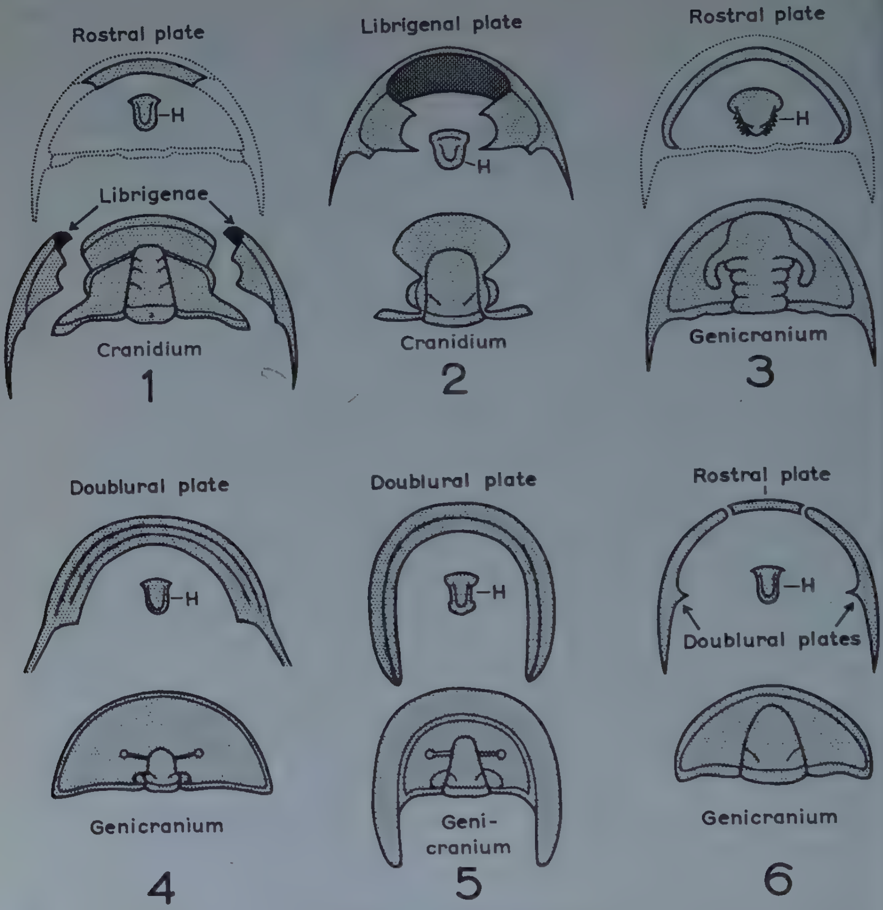

Fig. 1. Examples of cephalic plates in trilobites with facial sutures (1 — *Ptychoparia*, 2 — *Levisella*), in an olenellid (3) and in forms with a wholly or predominantly marginal suture (4 — *Harpides*, 5 — harpid, 6 — *Conocoryphe*). H = hypostome.

the hypostome, and a main plate comprising the dorsal part of the cephalon and a doublural part. Rather than to coin a new term for this main plate, I suggest to use the term genicranium in this case, too. This is consistent with the usage of the term cranidium, which may include a part of the doublure (many forms) or may not (*e.g.* asaphids).

No hypostome has as yet been described in agnostids, and the cephalon apparently consists of a single plate, which may of course be called the cephalon. However, if agnostids did have hypostome, the main plate should rather be termed the genicranium.

For the fused rostral plate and hypostome I propose the term *rostri-hypostomal plate*, and for the united librigenae the term *librigenal plate.*

FAMILY OLENIDAE BURMEISTER, 1843

Classification: — I have discussed the classification of the olenids in a previous paper (HENNINGSMOEN, 1957), but could not then make use of a paper published in the same year by HARRINGTON and LEANZA, who erected a family Hypermecaspididae. As discussed below (p. 161), I prefer to regard the Hypermecaspidinae as a subfamily of the Olenidae.

Origin and relationships: — I have earlier (1957, p. 30, and text fig. 3 on p. 23) suggested that the Olenidae might have developed from the Andrarinidae. At that time I thought that the aphelaspids might be included in the Andrarinidae, but they are now assigned to the family Pterocephaliidae LOCHMAN, 1956, which has been suggested to have developed from the Olenidae (LOCHMAN-BALK & WILSON, 1958, text fig. 10 on p. 330). I agree that the Pterocephaliidae may be rather closely related to the olenids, but believe that they either developed from the same ancestral group (as suggested in 1957, p. 31) or that the olenids developed from the aphelaspids through forms like *Olenus alpha* HENNINGSMOEN, 1957.

It has been suggested that *Parabolinoides* and related genera, now assigned to the Parabolinoididae LOCHMAN, 1956, are derived from the Olenidae (WILSON, 1954, p. 265; WILSON, 1957, p. 331; see also LOCHMAN-BALK & WILSON, 1958, text fig. 10). However, I still do not think this is the case (cf. HENNINGSMOEN, 1957, p. 22). Most probably the Parabolinoididae may be traced back to some earlier North American genera (and be related to e.g. *Elvinia*, cf. HENNINGSMOEN, 1957, p. 28). If so, the fact that the Parabolinoididae are characteristic of the *Conaspis* zone in the so-called cratonic realm, whereas there is little evidence of the *Conaspis* fauna in the miogeosynclinal areas, would no longer be puzzling. As stated by LOCHMAN-BALK & WILSON (1958, p. 338), this would be an enigma if the Parabolinoididae were derived from olenids, characteristic of the so-called extracratonic realm.

It seems probable to me that the Olenidae, Pterocephaliidae, Parabolinoididae, Idahoiidae, Elviniidae, and genera like *Irvingella*, *Dunderbergia*, and *Conaspis* together with its probable forerunner *Comanchia* are all related and might be grouped in a higher taxonomic unit. This unit could also embrace derived families, like the Ptychaspididae, developed from *Conaspis* through *Eoptychaspis* (cf. NELSON, 1951, p. 777), the Saukiidae, whether they developed from the Ptychaspididae (cf. RAASCH, 1952, p. 149), from *Conaspis* (cf. LOCHMAN, 1956, p. 451), or from some other related group, furthermore the Dikelocephalidae, whether they developed from a conaspid stock or from the Idahoiidae (cf. LOCHMAN-BALK & WILSON, 1958, p. 335, footnote 2), and probably also the Hungaiidae (cf. below). It is possible that the *Loganellus*-group and the Remopleurididae likewise belong here. This group of families apparently is too large and variegated to be regarded as a superfamily, and I suggest to accomodate them in a suborder Olenina of the order Ptychopariida. A suborder Olenina was recognized by HARRINGTON & LEANZA, 1957, but they attributed it to HUPÉ, 1953, who, however, only erected a superfamily Olenoidæ. The suborder Olenina might be attributed to SWINNERTON, 1915, who proposed a section Olenina of his suborder Conocoryphida. No doubt Middle Cambrian families should be included in the Olenina as well, perhaps *e.g.* the Andrarinidae and Anomocaridae. The Asaphidae and Ceratopygidae of the suborder Asaphina may be related to the Olenina (cf. HENNINGSMOEN, 1957, p. 28).

The relationships have not been traced for all the families suggested above to belong to the Olenina, but I believe that their possible relationships are worth considering, for which reason the above was written.

SUBFAMILY PELTURINAE HAWLE & CORDA, 1847

Genus *Peltocare* HENNINGSMOEN, 1957

Type species: — *Acerocare norvegicum* MOBERG & MÖLLER, 1898, by original designation.

Peltocare modestum n. sp.

Pl. 1, figs. 9-10.

Name: — From Latin *modestus*, modest.

Holotype: — A cranidium, P.M.O. no. 69565, pl. 1, fig. 9.

Material: — In addition to the holotype, only two more cranida, P.M.O. nos. 69566 and S 1932a.

Occurrence: — Ceratopyge Limestone (3a γ), near base of the thick, very fossiliferous bed, Bjørkåsholmen, Asker, Norway.

Diagnosis: — Differs from the type species in having somewhat narrower fixigenae, a more rounded front to the glabella, a more rounded margin to the cranidium between the eyes, and possibly in being a smaller species.

Description: — Cranidium about 1.5 times as wide as long, rather convex. Glabella pelturoid, well rounded in front and with 2 pairs (S1, S2) of faint furrows. Fixigenae pelturoid, only slightly more than half as wide as occipital ring. Eyes small, slightly behind the front of glabella, and rather close to the glabella. Preglabellar field short (sag.). Anterior margin rather convex, subparallel to the front of glabella. Posterior, occipital, and dorsal furrows distinct. Surface smooth. The length of the holotype cranidium is 3,0 mm and of another cranidium 2.7 mm. Other parts unknown.

Affinities: — *Peltocare modestum* n. sp. is no doubt close to the type species, but differs in some characters mentioned in the diagnosis, also when compared with cranidia of *P. norvegicum* of the same size. If the three known cranidia of *P. modestum* belong to adult specimens, *P. modestum* is considerably smaller than *P. norvegicum*, whose cranidium may reach a length of at least 15 mm. The rounded anterior margin of the cranidium of *P. modestum* seems to distinguish it from all other species assigned to *Peltocare*, although *P. olenoides* (SALTER, 1866) from the Upper Tremadoc of Wales is too poorly known to allow a comparison of this character.

FAMILY UNCERTAIN

Genus *Tropidopyge* HARRINGTON & KAY, 1951

Type species: — *Dicellocephalus Bröggeri* MOBERG & SEGERBERG, 1906, by original designation.

Tropidopyge broeggeri (MOBERG & SEGERBERG, 1906)

Pl. 1, figs. 5—7.

1906 *Dicellocephalus Bröggeri* n. sp. — MOBERG & SEGERBERG, p. 87, pl. V, figs. 7—8. (Descr., figs. of 2 pygidia.)

1951 *Dikelocephalus Bröggeri* MOBERG & SEGERBERG — HARRINGTON & KAY, p. 663. (Selected as type species of *Tropidopyge*.)

1957 *Tropidopyge broeggeri* (MOBERG & SEGERBERG) — HARRINGTON & LEANZA, p. 120. (Suggested to belong to the Hypermecaspididae.)

Lectotype: — The pygidium figured by MOBERG & SEGERBERG, 1906, pl. V, fig. 7, from the Ceratopyge Limestone at Ottenby in Öland, Sweden, selected by HARRINGTON & KAY, 1951.

Norwegian material: — Two incomplete pygidia, P.M.O. nos. 756a and 69567, the latter with its counterpiece (no. 69568) preserved.

Occurrence: — Ceratopyge Limestone (3a*y*), Norway: Stensberggata in Oslo; Sjøstrand in Asker (in a dark limestone nodule from the base of the Ceratopyge Limestone). Ceratopyge Limestone, Sweden.

Description of Norwegian material: — The larger (pl. 1, fig. 7) of the two pygidia is 14 mm long and shows no postaxial ridge. In this and in other features it agrees well with the larger pygidium figured by MOBERG & SEGERBERG (1906, pl. V, fig. 7), which is furthermore of about the same size. The smaller Norwegian pygidium (pl. 1, figs. 5—6) has a postaxial ridge and is very similar, both in morphology and shape, to the smaller pygidium figured by MOBERG & SEGERBERG (1906, pl. V, fig. 8). One might suspect that the smaller and larger pygidia represented two species, but from the original description it appears that intermediate sizes are known. Characteristic of the pygidium of *Tropidopyge broeggeri* are its subelliptical outline, short axis reaching only slightly more than halfway to the posterior border, short pleural furrows, wide doublure, and distinct terrace lines both on its dorsal surface and the doublure.

Affinities: — The only other species referred to *Tropidopyge*, *T. stenorhachis* HARRINGTON & KAY, 1951, from the Tremadocian of Colombia, resembles, but may not necessarily be congeneric with *T. broeggeri*.

The genus *Tropidopyge* was originally assigned to the family Dikelocephalidae by HARRINGTON & KAY (1951, p. 663), but was believed probably to belong to the family Hypermecaspididae HARRINGTON & LEANZA, 1957, by those who established the family. The only other genus included in this family is *Hypermecaspis* itself. As pointed out by HARRINGTON & LEANZA (1957, p. 120), *Hypermecaspis* seems most closely allied to *Parabolinella* BRØGGER, 1882, from which genus I believe it developed. The main reason for separating the Hypermecaspididae from the Olenidae appears to be the trend in *Hypermecaspis* towards an enlarged pygidium, whereas the pygidium is always small in the Olenidae, except for the aberrant pygidium of *Ctenopyge* *pecten*. However, the genus *Hypermecaspis* embraces species with a small pygidium (*H. armata*) of the olenid type as well as species with a relatively large pygidium (*H. inermis*), and I prefer to regard the hypermecaspidids as constituting a subfamily of the family Olenidae.

As observed by HARRINGTON & LEANZA (1951), the pygidium of *Tropidopyge* resembles that of *Hypermecaspis*. However, it also resembles the pygidium of several other trilobites, *e.g.* the pygidium of "*Pterocephalina*" *utahensis* figured by RESSER (1942, pl. 15, figs. 7–11) or, as pointed out by MOBERG & SEGERBERG (1906, p. 88), *Dikelocephalus pepinensis* OWEN (now assigned to *Saukiella*). SDZUY (1955, p. 37) suggested that *Tropidopyge* might belong to the subfamily Paracoosiinae KOBAYASHI, 1950, in which subfamily the genus *Pterocephalina* RESSER was included. MOBERG & SEGERBERG (1906, p. 90) cautioned that we do not know for certain whether the cranidium assigned to *Dikelokephalina discraeura* really belongs to this species, and stated that there is a possibility that it belongs to *Tropidopyge broeggeri*. This is not very likely, but all in all, I believe it is best not to assign *Tropidopyge* to any family for the time being.

If *Tropidopyge broeggeri* is related to *Hypermecaspis*, there is a possibility that it is conspecific with *Hypermecaspis* ("*Parabolinella*") *rugosa*, in which case *Hypermecaspis* might be considered as a junior synonym of *Tropidopyge*. KOBAYASHI (1951, p. 13) suggested that *Dikelokephalus broeggeri* might be congeneric with the type species of *Hagiorites* KOBAYASHI, 1951, based on an imperfect pygidium. If so, *Hagiorites* would become a junior synonym of *Tropidopyge*.

While this paper was in print, RASETTI has published an illustration of a pygidium of *Richardsonella subcristata* RASETTI (Jour. Paleont., vol. 33, pl. 55, fig. 18), which is very similar to that of *Tropidopyge broeggeri*.

FAMILY HUNGAIIDAE RAYMOND, 1924

SUBFAMILY DIKELOKEPHALININAE KOBAYASHI, 1936

Remarks: — KOBAYASHI (1936) regarded the Dikelokephalininae as a subfamily of the family Dikelocephalidae MILLER, 1890, and was followed in this *i.a.* by RICHTER & RICHTER (1954). RAW (1949, p. 514) suggested that *Dikelokephalina* might be a near relation of *Hungaia*, and HUPÉ (1953) considered the Dikelokephalininae as a subfamily of the Hungaiidae. SDZUY (1955) regarded the Dikelokephalininae as a subfamily of the Anomocaridae POULSEN, 1927, together with the Anomocarinae, Hungaiinae, and Paracoosiinae KOBAYASHI, 1950, whereas HARRINGTON & LEANZA (1957) considered the Dikelokephalinidae as a separate family.

I agree with the later views that the similarities between the dikelokephalinids and dikelocephalids may be an example of homoeomorphy, rather than indicating close relationships. Thus the glabellar patterns of the two groups are rather different. I believe RAW and HUPÉ are right in assuming relationships between *Hungaia* and the dikelokephalinids. Although the glabella of *Hungaia* apparently is the more advanced, the cephala of the two groups have much in common, including the alar-like inflations, and so have their pygidia. The relationships between the Anomocaridae and Hungaiidae are as yet uncertain, and I prefer to follow HUPÉ in regarding the Dikelokephalininae as forming a subfamily of the Hungaiidae.

There can be little doubt that *Dikelokephalina* is closely related to *Asaphopsis* MANSUY, 1920, and KOBAYASHI (1936, p. 175) pointed out that they might be members of a continous series of gradual transitions. *Asaphopsis intermedia*, described from the Llanvirnian of Argentina by HARRINGTON & LEANZA (1957, p. 191), is interesting in having 2 pairs of pygidial spines, the outer pair being situated as the pair in *Asaphopsis*, the inner pair as the pair in *Dikelokephalina*. *Hungioides* KOBAYASHI, 1936, apparently is related, and these three genera were united in the subfamily Dikelokephalininae by KOBAYASHI (1936), who later added *Dactylocephalus* and *Hagiorites*, both KOBAYASHI, 1951. Another genus, *Asaphopsoides*, was added by HUPÉ, 1953.

An interesting form was redescribed by SDZUY (1955, p. 37) as *Pterocephalina* (*Leimitzia*) *bavarica* (BARRANDE, 1868), and made the type species of the new subgenus *Leimitzia*. The genus *Pterocephalina* RESSER, 1938, has since (PALMER, 1956) been shown to be a junior synonym of *Litocephalus* RESSER, 1937. *Leimitzia* can hardly be regarded as a subgenus of *Litocephalus* as now defined, and I regard *Leimitzia* as a separate genus. *Leimitzia* agrees with the Dikelokephalininae in the general structure of the cranidium, also in having alar-like inflations, and its pygidium resembles those of the Dikelokephalininae, although it apparently does not carry any spines. I prefer to assign *Leimitzia* to the Dikelokephalininae, rather than to the

Pterocephaliidae, to which *i.a. Litocephalus* is now referred. It is, however, quite possible that the similarities between these two groups indicate relationships. As mentioned by SDZUY (1955), another probable member of *Leimitzia* is *Dikelocephalus celticus* SALTER, 1866, based on pygidia, with the associated and probably conspecific *D. discoidalis* SALTER, 1866, based on parts of the cephalon. *Leimitzia celtica* is known to occur in the Upper Cambrian zone of *Parabolina spinulosa* (cf. STUBBLEFIELD, 1951, p. 56), and would then be the earliest known member of the Dikelokephalininae, the next earlier being *Leimitzia bavarica* from the Lower Tremadocian. The similarities between *Leimitzia* and the *Pterocephaliidae* thus become especially significant.

Genus *Dikelokephalina* BRØGGER, 1896

Type species: — *Centropleura? dicraeura* ANGELIN, 1854, designated by VOGDES, 1925.

Dikelokephalina dicraeura (ANGELIN, 1854).
Pl. 1, figs 1—4.

1854 *Centropleura? dicræura.* n. sp. — ANGELIN, p. 88, pl. XLI, fig. 9. (Short diagn., fig. of pygidium).
1869 *Dikelocephalus dicræura* ANG. — LINNARSSON, p. 71 (Recorded.)
1882 *Dicelocephalus dicræurus,* ANG. — BRÖGGER, p. 126. (Remarks.)
1896 *Dikelokephalina dicræura,* ANG. — BRÖGGER, pp. 177—179 (reprint, pp. 14—16), text fig. 4. (Compared with similar forms, fig. of pygidium.)
1900 *Dicellocephalina dicræura* ANG. sp. — MOBERG, p. 534, pl. 14, fig. 1 (Descr. and fig. of cranidium assigned to this species).
1906 *Dicellocephalina dicræura* ANGELIN sp. — MOBERG & SEGERBERG, p. 90, pl. V, figs. 13—14, 12? (Remarks. Figs. of 2 pygidia, and a cranidium assigned to this species.)
1919 *Dikelocephalina dicræura* (ANGELIN) — LAKE, pp. 117—120. (Compared with *D. furca.*)
1925 *Dikelokephalina dicraeura* ANGELIN — VOGDES, p. 98. (Listed as genotype of *Dikelokephalina.*)
1956 *Dikelokephalina dicraeura* (ANGELIN) — TJERNVIK, p. 278. (Listed).

Type data: — Holotype (by monotypy) is the pygidium figured by ANGELIN (1854, pl. XLI, fig. 9) from Ceratopyge Beds ("Ceratopygarum"), most probably Ceratopyge Limestone, at Gamlebyen ("Opslo") in Oslo, Norway.

Norwegian material. — Four more or less fragmentary pygidia are preserved in Paleontologisk Museum, Oslo, *viz.* nos. 847, S 1175, 69572 (and counterpiece 69573), and 69575 (and counterpiece 69574).

Occurrence: — Ceratopyge Limestone (3aγ), Norway: Road cut W. of Slemmestad in Røyken; Bjørkåsholmen in Asker; Bygdøy Sjøbad, Trefoldighetskirken, and Gamlebyen in Oslo. Ceratopyge Limestone, Sweden.

Description: — The outline of the pygidium is not evenly curved. The anterior margins are short and almost straight, the antero-lateral corners well rounded, shoulder-like, the lateral margins almost straight, diverging somewhat to the rear, thus forming an obtuse and well rounded angle with the posterior margin, which is evenly convex, except for the two posterior spines and the intervening concave margin. The spines are flat and broad-based, but taper rapidly, and, as seen in plate 1, fig. 2, continue as round and slender spines. The axis has 6 distinct rings and a 7th poorly separated from the endlobe, which merges into a short and triangular postaxial ridge, which does not reach the posterior margin. The pleural furrows are wide and distinct inside the doublure, narrower and less well defined further out, and do not completely cross the brim. The upper surface of the pygidium is ornamented with rather close packed terrace lines, and the wide doublure with less closely spaced ones. The largest pygidium present is 30 mm long, spines not included.

Affinities: — *Dikelokephalina dicraeura* closely resembles *D. furca* (SALTER, 1866) from the Upper Tremadoc of Wales, as observed by LAKE (1919, p. 119), who pointed out that the outline of the pygidium of *D. furca* differed from that of *D. dicraeura*, as known then. It is now known that they agree even in this character. The cranidium assigned to *D. dicraeura* resembles that assigned to *D. furca*, and the difference noted by LAKE may be due to different preservation, as suspected by him. It is thus possible that the two forms are conspecific, but better material is needed to decide this.

FAMILY HARPIDIDAE[1] RAW, 1949

(*nom. transl.* WHITTINGTON, 1950, *ex* Harpidinae RAW, 1949).

Relationships within the family: — *Harpides* apparently is related to the two North American genera *Loganopeltis* RASETTI, 1943, and

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[1] The family name based on *Harpides* (*gen. sing. Harpidae*, stem *Harpid-*) is Harpididae, not Harpidedidae, as suggested by PRANTL & PŘIBYL (1955).

Loganopeltoides RASETTI, 1945, and it was for these three genera that RAW (1949) suggested the subfamily Harpidinae (of the family Harpidae[1]). WHITTINGTON later (1950a, 1950b) suggested to regard this group as a separate family. It seems unnecessary to split it into two families, Loganopeltoidae (*Loganopeltis*, *Loganopeltoides*) and Harpididae (*Harpides*, *Dictyocephalites*), as proposed by HUPÉ (1953), although *Loganopeltoides* is a proparian form (this was doubted by RAW, 1949, who considered it hypoparian), and both it and *Loganopeltis* (hypoparian) differ from *Harpides i.a.* in having a smaller brim. So far it has been possible to compare only the genicrania of these genera, but the great resemblance of these strongly suggest that they should be placed in the same family. It should be pointed out that even *Loganopeltis* shows smooth alar-like areas, so distinct in *Harpides*. Besides similar genae, all three genera have in common the type of glabella, the granulose surface ornamentation, and the radiating ridges.

Relationships to other families: — WHITTINGTON (1950b, p. 302) lists some of the differences between *Harpides* and harpids, *viz.* that the bilaminar border in *Harpides* was without the prolongations and the characteristic division into the inner, steep cheek roll and outer brim, and that it may have lacked the structure of opposed pits. We still do not know whether *Harpides* had opposed pits in the bilaminar border, but if not, they may easily be imagined to have developed from the pit-like depressions between the finer anostomosing ridges. *Harpides* and harpids have so many features in common that they most probably are closely related. Some of these features are the marginal suture, the bilaminar border, the alae, the pits, and also the preglabellar boss and the radiating ridges, so well developed in *Harpides* and present in many harpids. Furthermore, the glabella in *Selenoharpes* is very similar to that of *Harpides*. All this does not necessarily mean that the harpids developed from *Harpides*, which is perhaps less likely now that the earliest known harpid, *Australoharpes depressus* HARRINGTON & LEANZA, 1957, is reported from the same Lower Tremadocian zone as the earliest known species of *Harpides*, *H. neogaeus*. On the whole, the harpidids and harpids have so much in common, that the harpidids might just as well be regarded as forming a primitive subfamily of the Harpidae.

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[1] The genus *Harpes* was erected by GOLDFUSS (1943, p. 584), who explicitly stated that it was named after the cyclops *Harpes*. Being a personal name, *gen. sing.* is *Harpis*, and the stem *Harp-* (3rd. declination). The family name based on *Harpes* thus becomes Harpidae, as pointed out by RICHTER in 1943, and not Harpedidae as maintained by PRANTL & PŘIBYL (1955).

RAW (1949, p. 514) remarked that the Harpididae seem to be related to the Trinucleidae as well. In spite of similarities in the bilaminar border, there are still important differences, like the shape and furrows of the glabella, and it seems difficult at present to trace any relationships (cf. WHITTINGTON, 1950a, p. 6). The Entomaspididae show resemblances both to the Harpididae and Trinucleidae, as pointed out by RASETTI (1952, p. 801), but as cautioned by him, only the discovery of intermediate forms between the Trinucleidae and Entomaspididae could substantiate the hypothesis of relationship between the two families. In many features (*e.g.* the glabella) the Entomaspididae resemble much more the Harpididae than the Trinucleidae.

As to the ancestral stock from which the Harpididae developed. *Loganopeltoides* indicates that they had facial sutures. The peculiar type of proparian sutures in this genus probably evolved from more "normal" facial sutures, quite possibly opisthoparian sutures. It is interesting to notice the many similarities in the more central parts of the cephalon in *Harpides* and the cranidium assigned to *Dikelophalina dicraeura* (as figured by MOBERG & SEGERBERG, 1906, pl. V, fig. 12). Thus the shape of the glabella and glabellar furrows are much the same, both have small eyes and distinct eye ridges, as well as alae. As furthermore the pygidium of the related genus *Hungaia* has much in common with the pygidium of *Loganopeltoides*, one should consider the possibility that the Harpididae and Hungaiidae developed from the same stock. It is interesting that *Hungaia*, too, has a wide brim with well developed radiating ridges.

Genus *Harpides* BEYRICH, 1846

Type species: — *Harpides hospes* BEYRICH, 1846, by monotypy.

Harpides rugosus (SARS & BOECK, 1838)

Pl. 2, figs. 5—11.

1838 *Trilobites rugosus* Ss. & Bk. Mscr. — BOECK, p. 143. (Short descr.).
1854 *Harpides rugosus* SARS & BOECK — ANGELIN, p. 87, pl. XLI, figs. 7–7a. (Short diagn., fig. of genicranium.)
1869 *Harpides rugosus* SARS & BOECK — LINNARSSON, p. 67. (Recorded.)

1882 *Harpides rugosus*, Sars & Boeck — Brøgger, p. 127. (Remarks.)
1906 *Harpides rugosus* Sars & Boeck — Moberg & Segerberg, pl. 85, p. V fig. 1. (Remarks, fig. of fragmentary genicranium).
1906 *Harpides rugosus* Sars & Boeck — von Post, p. 476, pl. 13, figs. 3–5. Listed. Figs. of genicrania, including Angelin's original.)
1940 *Harpides rugosus* (Sars & Boeck Ms) — Stormer, p. 146, pl. 1, figs. 14–15. (Cites Boeck's original descr. Figs. of 2 fragmentary genicrania, one selected as lectotype.)
1949 *Harpides rugosus* (Sars and Boeck) — Raw, p. 511. (Remarks.)
1950 *Harpides rugosus* (Sars and Boeck) — Whittington, 1950b, p. 302. (Remarks.)
1956 *Harpides rugosus* (Sars & Boeck) — Tjernvik, p. 268. (Remarks.)

Lectotype: — A fragmentary genicranium (P.M.O. no. 20053) from the Ceratopyge Limestone, Oslo, Norway, selected by Størmer, 1940.

Norwegian material: — Ten more or less fragmentary genicrania (P.M.O. nos. 846, 1176, S 3037, 20052, 20053, 35935, 60321a, 69581, 69582, 69583) and two fragmentary doublural plates (P.M.O. nos. 1290, 56024a).

Occurrence: — Ceratopyge Limestone (3a*y*), Norway: Slemmestad and Ramtonholmen in Røyken; Bjørkåsholmen in Asker; Vekkerø and Rikshospitalet in Oslo; Steinsodden in Ringsaker. Ceratopyge Limestone, Sweden.

Description: — Four fragmentary genicrania (pl. 2, figs. 5, 7—9) together show the morphology of practically the whole genicranium. The differences in length-width ratio may be due to distortion.

The glabella is truncated in front and tapers forwards, but not evenly, since L2 is somewhat withdrawn. Occipital furrow with a distinct forward curve in front of the occipital node. S1 distinct, straigth, oblique backwards, reaching one-third across glabella. S2 faint, short, transverse. S3 very faint and short, oblique forwards. Gena semicircular with upsloping, narrow border. Eyes small, on line with front of glabella, distance from glabella subequal to width of glabella in front. Eye ridges distinct, directed outwards and slightly forwards from glabella to eyes. Alae small but distinct. An elongate preglabellar boss is distinct but ill defined. Genal corners without spine. The surface of the test, including the ridges, is ornamented with granules of varying size, except for the smooth alae and the posterior and dorsal furrows. The internal mould of the test is likewise granulose, showing that the granules were hollow, as is also true of the radiating and anastomosing ridges of the gena.

The main radiating ridges are broader and further apart in a wide zone along the anterior border, but the space between them is occupied by a meshwork of very fine ridges. The depressions between them may be regarded as pits. They are arranged in groups (cf. WHITTINGTON, 1950b, p. 302), separated by the main ridges. Close to the upturned border, the main ridges branch into finer ridges, and no ridges cross the border. WHITTINGTON (1950b, p. 302) reports a faint impressed line along the inner margin of the zone with broad ridges in a specimen examined by him. As seen in a photograph (pl. 2, fig. 9) of part of this specimen, this line appears as a faint convex bend on the outer surface of the gena. Between this line and the border there are two other parallel lines, which appear as very faint concave bends on the gena.

Of special interest are two specimens (pl. 2, figs. 6, 10) which apparently represent parts of the cephalic doublure, until now not known in material assigned to *Harpides rugosus*, but assumed to have been present by RAW (1949) and WHITTINGTON (1950b). The specimens show the same type of ridges and granulation as the genicrania of *H. rugosus*, and no doubt belong to this species rather than to an undescribed trilobite. Although very fragmentary and not too well preserved, the one specimen luckily shows a part of the left genal corner (seen from below). The genal spine deviates slightly outwards from the course of the anterior border, as is the case in *H. grimmi* BARRANDE, 1872, and *H. neogaeus* HARRINGTON & LEANZA, 1957. The surface is traversed by radiating ridges, except for the anterior border, the genal corner, and the genal spine. They are, however, in both specimens intersected by a pair of rail-like ridges, subparallel to the anterior margin, and agreeing in position with two furrows seen on the impression of the doublure of a specimen of *H. grimmi*, figured by BARRANDE (1872, pl. 1, fig. 11) in a drawing reproduced by RAW (1949, pl. 83, fig. 3). The two ridges may be compared with a single ridge, the girder, in harpids and trinucleids. In some trinucleids there may likewise be more than one such concentric ridge, thus *Cryptolithus discors* has three (cf. STØRMER, 1930, pl. 6, fig. 2). The two ridges on the doublural plate assigned to *H. rugosus* apparently agree in position with the two faint concave bends on the outer surface of the gena of this species. It is likely that the 4 concentric depressions in *H. neogaeus* represent as many ridges on the doublure. The position of the inner margin of the doublure may be indicated by a convex bend on the surface of the gena, seen in the photograph of the specimen reproduced by HARRINGTON & LEANZA (1957, text fig. 103,1).

In analogy with the terminology for harpids, the two concentric ridges in *H. rugosus* may well be termed the girders, the doublure may be termed the inner lamella, and the outer zone of the genicranium, with broad ridges, may be termed the outer lamella. The doublural plates assigned to *H. rugosus* show that RAW (1949) and WHITTINGTON (1950b) were right in assuming that this species had a marginal suture and that the inner lamella bore the genal spines.

No hypostome, thorax, or pygidium have been assigned to *H. rugosus*.

Affinities: — The Bohemian species *Harpides grimmi* apparently is closely related to *H. rugosus*, or perhaps even a junior synonom of it, as suggested by RAW (1949, p. 511). The parts of the genicranium and doublural plate which can be compared are very similar, but, strangely enough, the impression of the inner lamella of *H. grimmi* does not seem to show any impressions of granulose or radiating ridges, as I have been able to ascertain also in a plaster cast of the specimen mentioned above. As furthermore the other parts of the exoskeleton can not be compared at present, it may be safest to regard the Bohemian and Scandinavian forms as two different species for the time being.

FAMILY OROMETOPIDAE HUPÉ, 1953.

Genus *Pagometopus* n. gen.

Name: — From Greek *pagos* (hill) and *metopon* (forehead), suggesting likeness to *Orometopus* (Greek *oros* = mountain).

Type species: — *Pagometopus gibbus* n. sp.

Diagnosis: — Cranidium resembling that of *Orometopus*, with a steep-sided and well elevated glabella, but with a strongly tapering and protruding frontal area.

Affinities: — Only the cranidium of *Pagometopus* is known, but it resembles the cranidium of *Orometopus* so much that it can hardly be doubted that the two genera are closely related. For this reason I place *Pagometopus* in the family Orometopidae. The well-developed preglabellar field of *Pagometopus* suggests relationships with the Hapalopleuridae HARRINGTON & LEANZA, 1957, of which *e.g. Araiopleura* HARRINGTON & LEANZA, 1957, has a somewhat similar glabella. However, the preocular parts of the facial sutures do not converge forwards in the same manner in the Hapalopleuridae as in *Pagometopus*. It is of interest that the pit at the outer end of the posterior border furrow in *Pagometopus* apparently is duplicated in certain members of the trinucleids, which are generally believed to be related to the orometopids (cf. illustration of *Trinucleus bronni* given by STØRMER, 1930, pl. 2, fig. 3).

Pagometopus gibbus n. gen., n. sp.

Pl. 2, figs. 1—4.

Name: — From Latin *gibbus*, humped, humpbacked.

Holotype: — A cranidium, P.M.O. no. 69577, pl. 2, fig. 1.

Material: — In addition to the holotype, only 2 more cranidia, P.M.O. nos. 69578 and 65979 (with counterpiece no. 69580).

Occurrence: — The thick, very fossiliferous bed of the Ceratopyge Limestone (3a*γ*), Bjørkåsholmen, Asker, Norway.

Diagnosis and description: — Cranidium about twice as wide as long, relatively flat, except for prominent glabella. Posterior margin rather straight, except for protruding V-shaped occipital part. Lateral margins tapering strongly forwards, incurved at eyes and more or less conspiciously on line with front of glabella. Anterior margin slightly convex. Glabella pyriform, widest anteriorly, bluntly rounded in front, well raised with steep sides, and with a keel-like bend in the middle of slightly more than the posterior half. S1-S3 represented by three faint pit-like impressions close to the dorsal furrow. Faint alar-like inflations are present. Posterior border furrow distinct, with a characteristic pit at its outer end. Eyes relatively small, distance from glabella about half the width of the postocular parts of the fixigenae, and distance from posterior margin about two fifths of the distance from a transversal line through tip of cranidium. Eye ridges distinct, directed outwards and slightly backwards. Preglabellar field about one fifth as long as glabella.

Outer surface of test appears smooth, inner side is granulose, giving the internal impression a pitted appearance. No glabellar or occipital spines (ascertained in a counterpiece).

The description is based on 3 cranidia, from 4.5 to 5.5 mm long. No other parts of the trilobite are known.

Affinities: — *Pagometopus gibbus* n. gen., n. sp. resembles species of *Orometopus*, but differs in the characters mentioned above. Furthermore, *P. gibbus* is larger than any described species of *Orometopus*.

FAMILY ALSATASPIDIDAE TURNER, 1940

Synonym: — Selenecemidae WHITTINGTON, 1952.

Genus *Falanaspis* TJERNVIK, 1956

Type species: — *Falanaspis aliena* TJERNVIK, 1956, by original designation.

Falanaspis aliena TJERNVIK, 1956

Pl. 1, fig. 8.

Type data: — Holotype is a genicranium, Palæontological Institute, Uppsala, no. Vg 389, from the zone of *Plesiomegalaspis armata* (lowermost Arenigian) at Stenbrottet, Västergötland, Sweden.

Norwegian material: — One fragmentary genicranium (P.M.O. no. S 1238) from the Ceratopyge Limestone (3a*y*) at Vekkerø in Oslo.

Remarks: — The Norwegian genicranium agrees well with the Swedish ones, also in having a glabellar node and faint, roundish S1 and S2. There seems to be no reason to describe it as a new form, although it occurs in the zone below that of the Swedish specimens.

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TREMADOCIAN TRILOBITES FOUND IN NORWAY AND THEIR STRATIGRAPHIC OCCURRENCE 2e = Dictyonema Shale 3a = Ceratopyge Beds * = not previously recorded from Norway	HORIZON ? = uncertain determ.						
	2e				3a		
	α	*β*	*γ*	*δ*	*α*	*β*	*γ*
AGNOSTIDAE							
**Geragnostus crassus* Tjernvik	.	.	.	.	.	.	+
— *sidenbladhi* (Linnarsson)	.	.	.	.	?	?	+
**Trinodus mobergi* Tjernvik	.	.	.	.	.	.	+
OLENIDAE							
Parabolina acanthura (Angelin)	+						
Bienvillia? wimani (Westergård)	.	+					
Bienvillia tetragonalis broeggeri Henningsmoen	.	.	.	.	+		
Parabolinella limitis Brøgger	.	.	.	.	.	+	
— *lata* Henningsmoen	.	.	.	.	.	.	+
Triarthrus angelini Linnarsson	.	.	.	.	.	+	+
Hypermecaspis rugosa (Brøgger)	.	.	.	.	.	.	+
Boeckaspis hirsuta (Brøgger)	+						
— *mobergi* (Wiman)	.	+					
Jujuyaspis keideli norvegica Henningsmoen	+						
— *angusta* Henningsmoen	.	+					
Peltocare norvegicum (Moberg & Möller)	.	.	.	.	+		
* — *modestum* n. sp.	.	.	.	.	.	.	+
Saltaspis sp.	.	.	.	.	.	.	+
**Plicatolina* sp.	.	.	.	.	.	+	
FAMILY UNCERTAIN							
**Tropidopyge broeggeri* (Moberg & Segerberg)	.	.	.	.	.	.	+
HUNGAIIDAE							
Dikelokephalina dicraeura (Angelin)	.	.	.	.	.	.	+
REMOPLEURIDIDAE							
Apatokephalus serratus (Boeck)	.	.	.	.	+	+	+
CERATOPYGIDAE							
Ceratopyge forficula forficula (Sars)	.	.	.	.	.	+	
— — *acicularis* (Sars & Boeck)	.	.	.	.	.	.	+
ASAPHIDAE							
**Promegalaspides (Borogothus) intactus* (Mob. & Seg.)	.	.	.	.	.	.	+
Platypeltoides incipiens (Brøgger)	.	.	.	.	+		
Niobe insignis Linnarsson	.	.	.	.	?	.	+
Niobella obsoleta (Linnarsson)	.	.	.	.	.	.	+
**Varvia longicauda* Tjernvik	.	.	.	.	.	.	+
NILEIDAE							
Nileus limbatus Brøgger	.	.	.	.	.	.	+
Symphysurus angustatus (Sars & Boeck)	.	.	.	.	.	.	+
SHUMARDIIDAE							
Shumardia pusilla (Sars)	.	.	.	.	?	+	+
EULOMIDAE							
Euloma ornatum Angelin	.	.	.	.	.	+	+
PLIOMERIDAE							
Parapilekia speciosa (Dalman)	.	.	.	.	.	.	+
Pliomeroides primigenus (Angelin)	.	.	.	.	.	.	+
OROMETOPIDAE							
Orometopus primigenius Størmer	.	.	.	.	+	.	
— *elatifrons* (Angelin)	.	.	.	.	.	.	+
**Pagometopus gibbus* n. gen., n. sp.	.	.	.	.	.	.	+
ALSATASPIDIDAE							
**Falanaspis aliena* Tjernvik	.	.	.	.	.	.	+
HARPIDIDAE							
Harpides rugosus (Sars & Boeck)	.	.	.	.	.	.	+

PLATES 1—2.

PLATE 1

The photographs are not retouched, but the specimens were coated with ammonium chloride before photographing. P.M.O. = Palaeontological Museum of the University in Oslo.

Dikelokephalina dicraeura (ANGELIN, 1854) — p. 163.

Fig. 1. × 1.45. Pygidium showing part of the doublure. P.M.O. no. 69572. 3a *y*, road cut west of Slemmestad, Røyken. Coll.: L. Størmer exc., 1953.

Fig. 2. × 1.45. Counterpiece of specimen in fig. 1, showing left pygidial spine. P.M.O. no. 69573.

Fig. 3. × 1.45. Pygidium. P.M.O. no. 69575. 3a*y*, Bjørkåsholmen, Asker. Coll.: G. Henningsmoen, 1958.

Fig. 4. × 1.45. Pygidium. P.M.O. no. 847. 3a*y*, Trefoldighetskirken, Oslo. Coll.: Rekdal, 1922.

Tropidopyge broeggeri (MOBERG & SEGERBERG, 1906) — p. 159.

Fig. 5. × 1.45. Pygidium. P.M.O. no. 69567. 3a*y* (dark limestone nodule), Sjøstrand, Asker. Coll.: G. Henningsmoen, 1958.

Fig. 6. × 1.45. Counterpiece of the specimen in fig. 5. P.M.O. no. 69568.

Fig. 7. × 1.45. Pygidium. An associated librigena disturbs the appearance of the left side. P.M.O. no. 756a. 3a*y*, Stensberggata, Oslo. Coll.: J. Kiær.

Falenaspis aliena TJERNVIK, 1956 — p. 171.

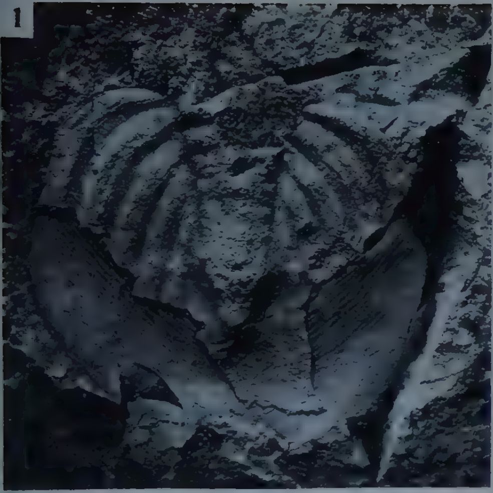
Fig. 8. × 4.1. Genicranium. P.M.O. no. S 1238. 3a*y*, Vekkerø, Oslo. Coll.: L. Størmer, 1919.

Peltocare modestum n. sp. — p. 158.

Fig. 9. × 4.5. Holotype cranidium. P.M.O. no. 69565. 3a*y*, Bjørkåsholmen, Asker. Coll.: G. Henningsmoen, 1958.

Fig. 10. × 4.5. Cranidium. P.M.O. no. 69566. 3a*y*, Bjørkåsholmen, Asker. Coll.: G. Henningsmoen, 1958.

PLATE 1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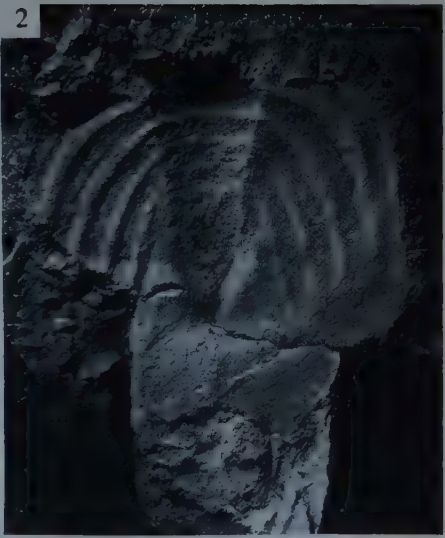
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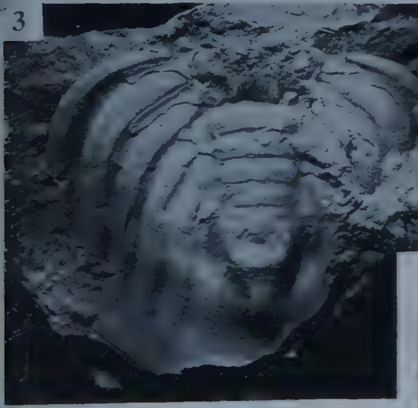
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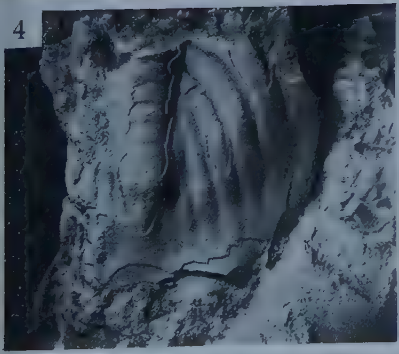
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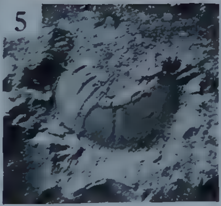
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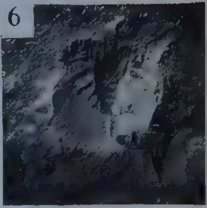
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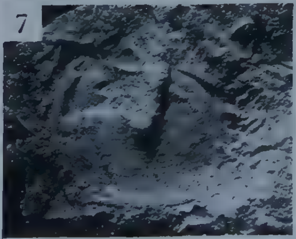
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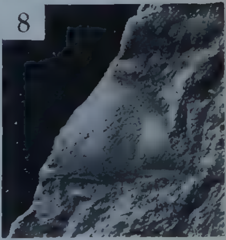
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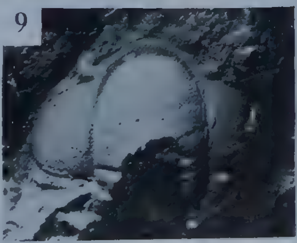
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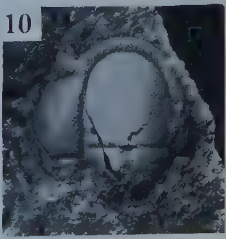
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PLATE 2

The photographs are not retouched, but the specimens were coated with ammonium chloride before photographing. P.M.O. — Palaeontological Museum of the University in Oslo.

Pagometopus gibbus n. sp., n. gen. — p. 170.

Fig. 1. × 4.5. Holotype cranidium. P.M.O. no. 69577. 3a*y*, Bjørkåsholmen, Asker. Coll.: G. Henningsmoen, 1958.

Fig. 2. × 4.5. Cranidium. P.M.O. no. 69578. 3a*y*, Bjørkåsholmen, Asker. Coll.: G. Henningsmoen, 1958.

Fig. 3. × 4.5. Cranidium. P. M. O. no. 69579. Left side exfoliated, showing the pitted surface of the internal impression. 3a*y*, Bjørkåsholmen, Asker. Coll.: G. Henningsmoen, 1958.

Fig. 4. × 4.5. Left side view of the specimen in fig. 3.

Harpides rugosus (Sars & Boeck, 1838) — p. 166.

Fig. 5. × 1.45. Genicranium. P.M.O. no. 69581. 3a*y*, Bjørkåsholmen, Asker. Coll.: F. Nikolaisen, 1958.

Fig. 6. × 1.45. Impression of part of doublural plate. P.M.O. no. 1290. 3a*y*, Bjørkåsholmen, Asker. Coll.: ? 1915.

Fig. 7. × 1.45. Genicranium showing left genal corner, P.M.O. no. 69582. 3a*y*, Bjørkåsholmen, Asker. Coll.: G. Henningsmoen, 1958.

Fig. 8. × 1.45. Genicranium. P.M.O. no. 69583. 3a*y*. Bjørkåsholmen, Asker. Coll.: G. Henningsmoen, 1958.

Fig. 9. × 2.2. Antero-latero-dorsal view of part of left genal area. Plaster mould of P.M.O. no. S 3037. 3a*y*, Vekkerø, Oslo. Coll.: P. Størmer, 1919.

Fig. 10. × 1.45. Left part of doublural plate, lower surface. P.M.O. no. 56024a. 3a*y*, Bjørkåsholmen, Asker. Coll. L. Størmer, exc., 1934.

Fig. 11. × 2.75. Detail of the specimen in fig. 8.

1

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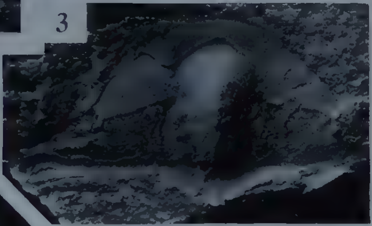

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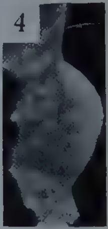

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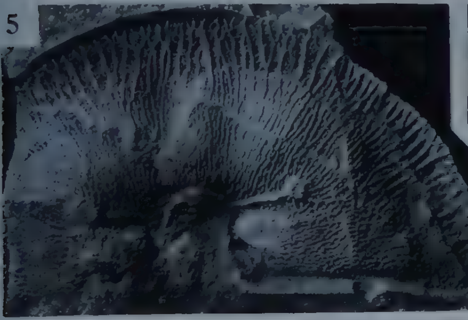

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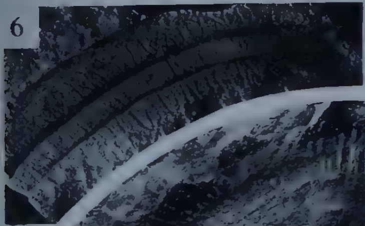

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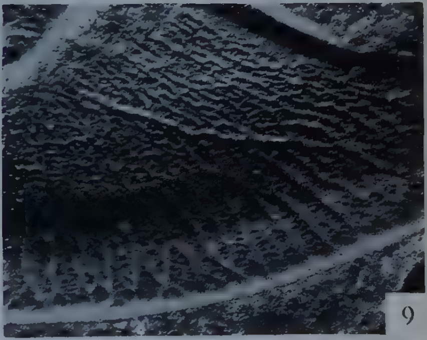

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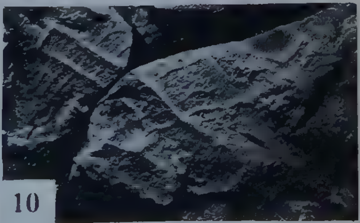

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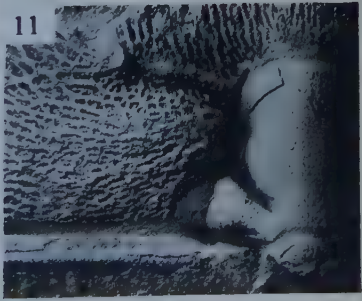

PEGMATITE VEINS AND THE SURROUNDING ROCKS

III. Structural control of small pegmatites in amphibolite, Rytterholmen, Kragerøfjord, Norway

BY

PAUL REITAN

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Abstract: A study was made of the structural position of 266 small, tabularly shaped pegmatites in an amphibolite in the pre-Cambrian Kongsberg—Bamble formation on the island Rytterholmen near Kragerø, southern Norway.

The pegmatites and the surrounding amphibolite are briefly described petrographically. Special attention is paid to the analysis of the structural position of the pegmatites in terms of shear-stress zones and tension directions which developed during the dynamo-thermal metamorphism of the area. The long axes of the pegmatites show a pronounced directional orientation in each of the five small areas examined in detail. This preferred orientation is related to zones of shear which are about perpendicular to the tension directions. The ratio length/width of the pegmatites is found to be roughly inversely correlated to the basicity of the surrounding amphibolite. The theory of origin by metamorphic differentiation is discussed and found to be the most satisfactory explanation of the genesis of the pegmatites. Basified zones in the amphibolite adjacent to the pegmatites and the difference between the An % in the surrounding amphibolite and the pegmatites substantiate this conclusion.

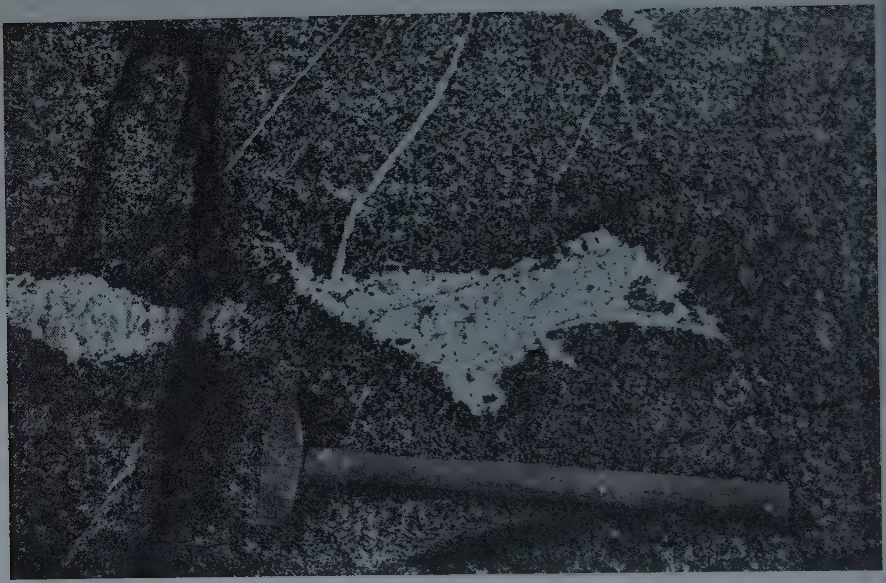


Fig. 1. Shows long, thin, conformable pegmatites (type 1) amputated by pegmatite body (type 3). Hammer is 48 ½ cm long. Photo at area A by P. Reitan.

Introduction

This paper is primarily a study of a particular type of small pegmatite veins found in the amphibolite exposed on the island Rytterholmen, located about 6 km almost due east of the town of Kragerø on the south coast of Norway.

Knowledge of the general geology and of special areas and characteristics of the Kragerø district are to be found in the works of Brøgger (4, 5), J. A. W. Bugge (6), Green (8), Hofseth (9), and Holtedahl (10). Especially useful for the area near Rytterholmen is the map by Brøgger (5).

Rytterholmen consists primarily of a gneissic amphibolite, which in zones is rather garnet (almandine-pyrope) rich, and a micaceous quartzite. The rocks have been thoroughly metamorphosed in the amphibolite facies, as indicated by the petrographic descriptions of the amphibolite on Rytterholmen (p. 178) and the descriptions of the petrography of the near-by islands by Brøgger (5) and Hofseth (9).

On Rytterholmen there are four different types of pegmatites in the amphibolite. The oldest pegmatite veins, type 1, which are cross-

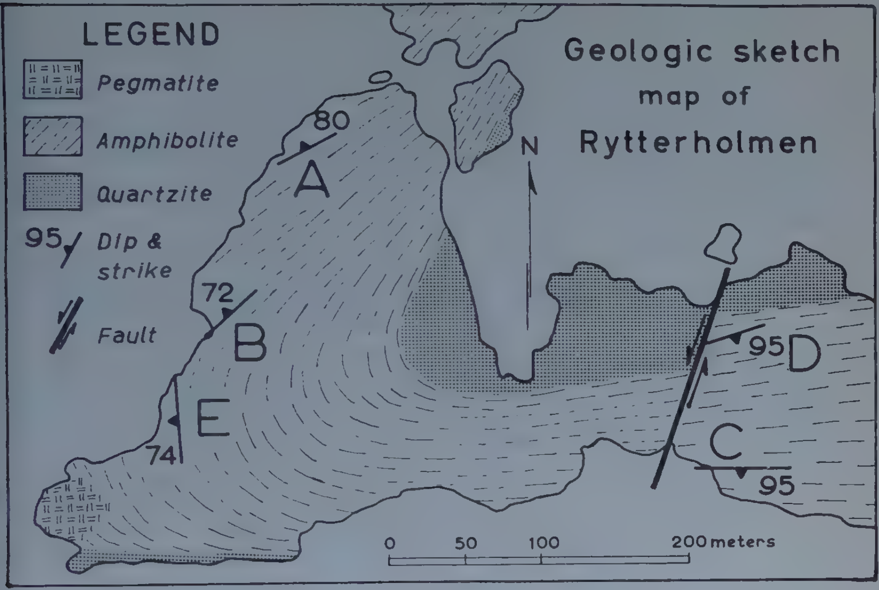


Fig. 2.

cut by all of the others, are thin, long, primarily plagioclase pegmatites which are conformable to the gneissic structure of the amphibolite (fig. 1). Two types were probably formed approximately contemporaneously: type 2 is an individual, massive, irregular pegmatite, found at the apex of the main anticline on Rytterholmen; type 3 consists of numerous, small pegmatites of sometimes slightly irregular, always elongate shape (tabular or pillow shaped in three dimensions). These have been the object of primary interest in this study. The youngest veins, type 4, which by virtue of their grain-size may be called pegmatites, are long, thin, crack-fillings composed either of quartz and a little calcite or calcite and dolomite. These veins crosscut all other structures and are clearly the youngest; they are probably a very late feature of the history of the area.

Petrography

Five small areas (fig. 2) of perfectly exposed rock were found on Rytterholmen. From these areas samples of the pegmatites of type 3 and the amphibolite were taken.

PEGMATITES

The pegmatites consist primarily of white plagioclase (sometimes with a very faint greenish tinge) and clear quartz, the former being predominant, with some biotite, occasionally a little hornblende, and rarely a little garnet. Garnet is present only in those zones where the amphibolite is garnet bearing.

The grain-size of the plagioclase and quartz is up to 2 cm in largest dimension, more normally nearer 1 cm. Biotite occurs in flakes up to 1 cm across, though generally near ½ cm. Hornblende and garnet, when present, are slightly finer-grained.

The composition of the plagioclase is quite constant; the most calcic plagioclase is An_{40} and the least calcic is An_{38} (table 1 and fig. 3).

Table 1

Composition of plagioclase in pegmatites[1]

Area	A	B	C	D	E
An % in plagioclase	38	38	40	39	38—39

[1] Determined by the refractive indices of cleavage flakes for the two orientations (001) and (010), the method of Tsuboi, and by the extinction angle in the plane ⊥ (001) and (010) and the maximum extinction angle in the zone || (010) using the universal stage.

AMPHIBOLITE

The amphibolite is mostly homogeneous, dark, medium-grained and shows reticular (network) fabric perpendicular to the linear structure. The essential minerals are hornblende and plagioclase; the most abundant mineral is always hornblende.

Hornblende occurs in rather elongate, idioblastic grains with ragged terminations, the orientation of which is responsible for the gneissic, linear structure of the rock. Hornblende shows pleochroism α = light olive-green, β = olive-green, and γ = dark green; absorption scheme $\gamma > \beta > \alpha$.

Plagioclase always constitutes more than 20 % of the rock. Its shape is irregular. It is usually slightly sericitized. At three of the areas from which samples were taken the composition of the plagioclase is An_{42}; at two areas the composition is An_{44} (table 2 and fig. 3). Albite and pericline twinning are very common.

Table 2

Modal analyses of thin-sections

Sample no.	Plagioclase				Horn-blende	Quartz	Bio-tite	Opa-ques	Apa-tite	Zir-con	Chlo-rite	Epi-dote	Cal-cite	Σ Dark[2]
	An[1]	Total	Albite	Anor-thite										
9A—RC1	44	20.8	11.7	9.1	62.5	6.9	0.9	7.1	1.5	*[3]	*	0.3	—[4]	70.5
9A—RE1	42	23.2	13.6	9.6	56.1	5.9	7.7	1.4	0.3	—	3.9	0.4	1.1	69.1
9A—RA1	42	27.8	16.1	11.7	57.8	8.7	1.0	3.7	0.3	*	0.4	—	0.3	62.9
9A—RD1	44	33.2	18.6	14.6	46.6	9.0	6.8	1.9	0.7	*	0.9	0.7	0.2	56.2
9A—RB1	42	40.6	23.6	17.0	48.2	6.3	2.2	2.4	0.3	*	*	—	—	52.8

[1] An content of the plagioclase determined by the extinction angle in the plane ⊥ (001) and (010) and the maximum extinction angle in the zone || (010) using the universal stage.

[2] The sum of hornblende, biotite, opaques, and chlorite (secondary after hornblende).

[3] The asterisk indicates that the mineral is present but in amounts less than 0.2 %.

[4] The dash indicates that the mineral was not observed in the thin-section.

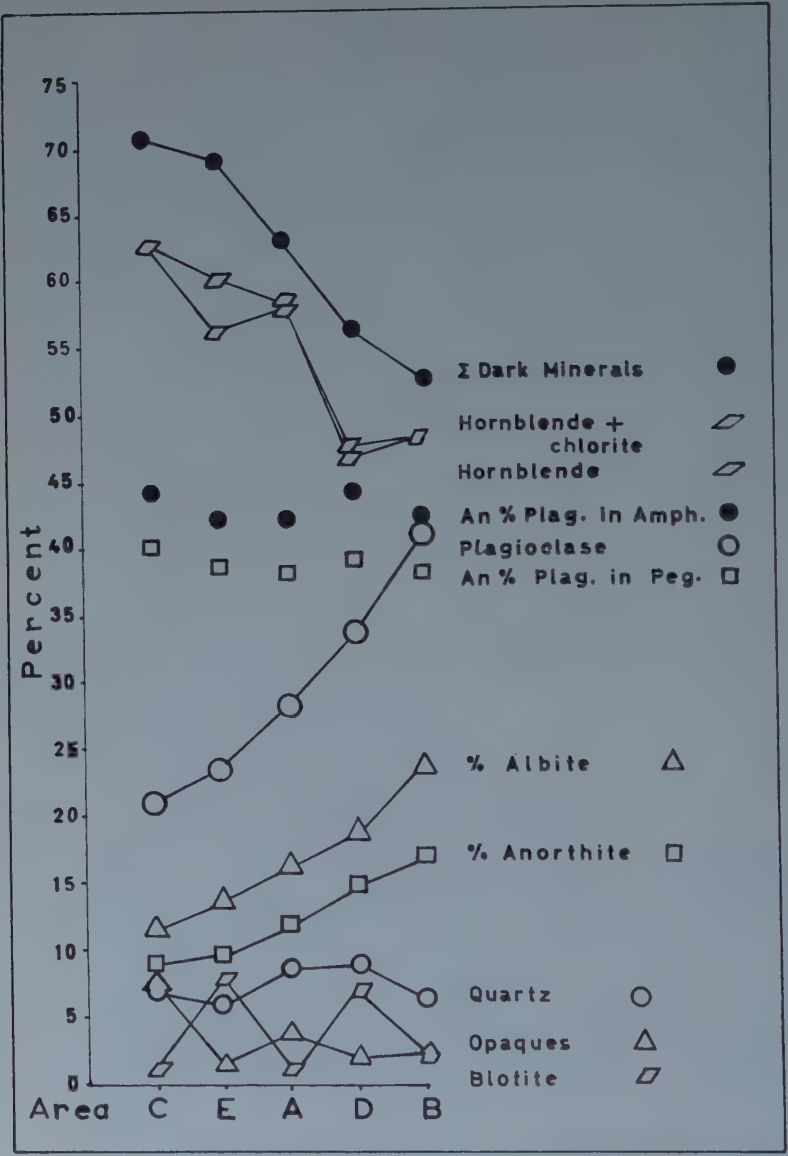


Fig. 3. Results of modal analyses. Total plagioclase is also separated into anorthite and albite. Anorthite percent in plagioclase from amphibolite and from pegmatites of type 3 shown by solid circles and open squares, respectively, unconnected by lines.

The accessory minerals always present are quartz, biotite, apatite, and opaques (pyrite, magnetite, and hematite). Zircon, epidote, pennine (secondary, primarily after hornblende), and calcite are frequently present and limonite is very rare.

Modal analyses of a typical sample of the amphibolite from each of the five areas were made; the results are shown in table 2 and fig. 3.

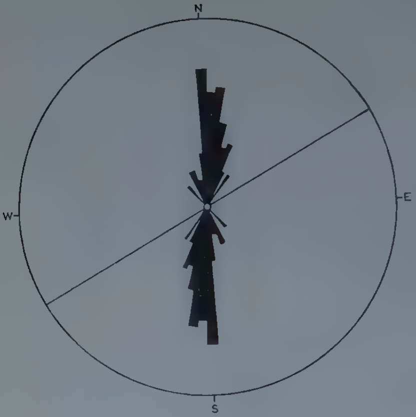
Structure

The primary structure on Rytterholmen is an anticline, the axis of which strikes N60E and plunges at about 75^g towards the SW. The axial plane dips very steeply towards the NW. The southern limb of the fold strikes N90—100E and dips 95^g S and the northern limb strikes about N50E and dips 80^g N (see map, fig. 2).

Table 3

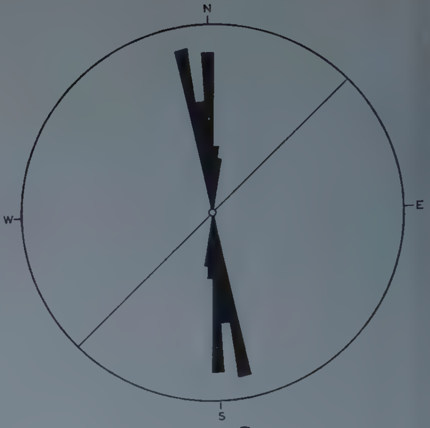
Orientation of the pegmatites at the five areas

Direction	Percent of the measurements				
	A	B	C	D	E
N 62½–68½ W				9	
57½–62½				21	
52½–57½				25	
47½–52½				15	
42½–47½	1		5	9	
37½–42½	0		19	6	
32½–37½	0		9	6	
27½–32½	0		29	9	7
22½–27½	2		5		13
17½–22½	2		19		48
12½–17½	1		14		20
7½–12½	1	38			7
2½– 7½	4	17			2½
N 2½ W–N 2½ E	26	35			2½
N 2½– 7½ E	18	6			
7½–12½	20	4			
12½–17½	10				
17½–22½	5				
22½–27½	6				
27½–32½	2				
32½–37½	0				
37½–42½	2				
Number of meas'm'ts.	85	52	21	33	75



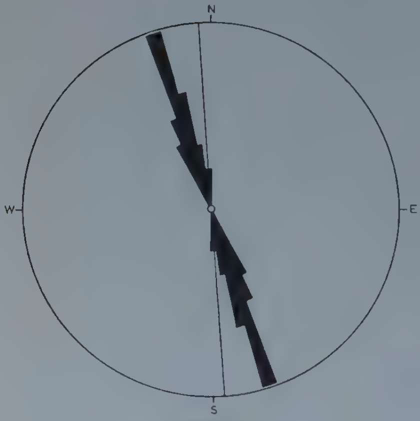
Area A

Fig. 4.



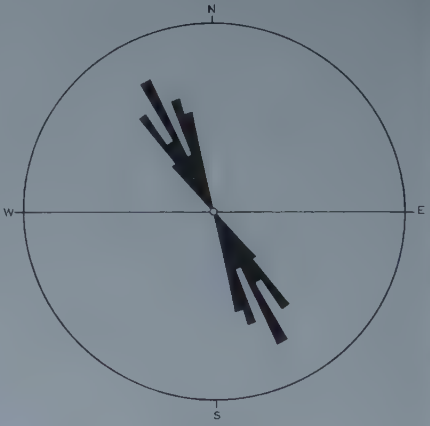
Area B

Fig. 5.



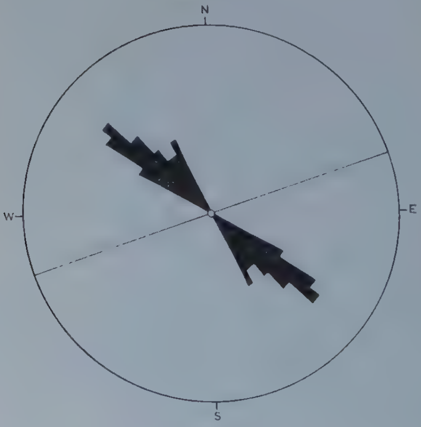
Area E

Fig. 6.



Area C

Fig. 7.



Area D

Fig. 8.

Fig. 4—8. Orientation of the long axes of the pegmatite bodies at areas A, B, E, C, and D. The area blackened within each 5^g segment of the circle is proportional to the percentage of measurements within that segment. The outer circle is at 49 %. The local strike is represented by the diameter drawn.

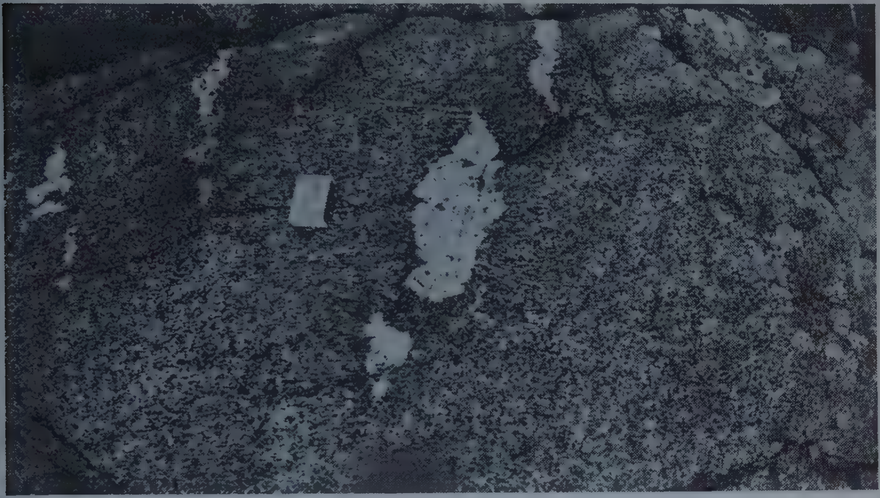


Fig. 9. Pegmatites of type 3 with change of strike of amphibolite and basified zones adjacent to them. The match box is 5 ½ cm long. Photo at area A by P. Reitan.

Within the amphibolite there are many small, elongate pegmatites. The strike direction of 266 of these distributed over the five selected areas was measured. It was possible to measure the dip of only about 25 % of these tabular or pillow shaped bodies. In all cases the dip was quite steep, varying between about 80^g and 95^g. Measurements were confined to virtually horizontal rock surfaces and therefore all of the measurements of the strike of the long axes of the pegmatites are used, even when the dip of the pegmatite was unobtainable. The results of these measurements are presented in table 3 and graphically in figs. 4 to 8.[1]

In these figures it can be seen that within each of the five areas there is a very pronounced preferred orientation of the long axes of the pegmatite bodies, though this direction is not the same in these areas nor is the angle between this direction and the strike of the amphibolite constant.

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[1] In these figures the blackened area, not the length, within each 5^g segment of the circle is proportional to the percentage of measurements which fell within that segment. Thus the first percent is accorded greater length than any thereafter, the length per percent diminishing as the width of the segmenth increases. The radius of the outer circle equals the length to which any 5^g segment would be blackened if 49 % of the measurements fell within it. The solid line transecting the circle represents the direction of the strike of the amphibolite.

The amphibolite is very clearly gneissic at areas A, C, and D, a little less strongly so at area B, and nearest the apex of the fold at area E the gneissic structure is discernable though with some difficulty. Near the pegmatite bodies there can be seen a change in the direction of the strike of the amphibolite (figs. 1, 9, and 10).

Genesis of the pegmatites

Only the pegmatites designated type 3 (p. 000) have been the object of special attention and therefore only the genesis of these pegmatites will be discussed in detail here. Of the others suffice it to note that the oldest pegmatites, those of type 1, occur as long, thin veins parallel to the strike, a structural position which corresponds to that along which one would expect concentric shear-planes due to elastic stress in an elastically folded layer (DE SITTER, 7, p. 74); type 2 occurs as a massive, irregular body at the apex of a fold, a site at which maximum tension and minimum pressure would be expected to prevail; and those of type 4 occur as fillings of apparently very late joints — also low pressure volumes.

The significance of the preferred orientations of the pegmatites of type 3 may not at first seem immediately self-evident and it is therefore intended to analyze the structural conditions which prevailed at the five areas in terms of tension and shear. The reader is asked to keep in mind that the pegmatites occur in an amphibolite band 150 m and more thick which is surrounded by a micaceous quartzite, and that the whole area was under the P, T conditions of the amphibolite facies during the dynamo-thermal metamorphism of the area. Such being the case the quartzite must be assumed to have been relatively plastic with respect to the amphibolite during deformation. Also, the pressure which caused the folding in the region concerned will be assumed to have been essentially perpendicular to the strike of the axial plane of the anticline revealed on Rytterholmen.

ANDERSEN (1 and 2) considered that the explanation to the openings in which new material crystallized, whether as perthite veins or as pegmatites in heterogeneous associations of minerals or rock bodies respectively, might be found in the unequal thermal contraction of the containing body as opposed to its surroundings. He also admitted as a serious possibility, and as often probable, that the openings might have been caused by the stresses set up by orogenic forces in zones of intensive folding. He did not advocate the latter because «... the meager data now available on the elastic properties of rocks do not give any definite proof of it.» (2, p. 49). It is intended here, in the light of present knowledge, to investigate the validity of the second possibility.

Area A is located on the northern limb of the fold. The regional pressure is not far from perpendicular to the boundaries of the amphibolite here and one would therefore expect two oblique shear directions to tend to develop, these forming acute angles with the boundaries of the band; the exact size of these angles would depend on the physical properties of the amphibolite. Area A is, however, much closer to the axis of the anticline revealed on Rytterholmen than to the next syncline (cf. BRØGGER's map, 5), and one may therefore expect rotational shear (cf. DE SITTER, 7, fig. 72, p. 100) which would in turn result in a tensional direction oblique to the strike of the amphibolite and directed towards the core of the anticline. The shearing between the amphibolite and the quartzite at the contacts and (what results in the same thing) the plastic flow of the quartzite will subject the amphibolite to a shear couple which will cause a tensional direction in the amphibolite — this also oblique to the strike of the amphibolite and directed towards the core of the anticline. Still one more direction of shear may be anticipated in a relatively competent layer surrounded by less competent layers as the folding begins to approach isoclinal folding. This shear direction is approximately parallel to the direction of regional pressure (fig. 11) and will be strongest at points furthest removed from the axes of the folds, i.e., where the limb of the fold strikes most directly across the direction of regional pressure.[1] One additional direction of tension is also to be expected, it is parallel to the strike — the direction of stretching as the layer is folded — tending to induce tension joints across the layer. This tendency is enhanced by the plastic flow of the surrounding quartzite. Tension due to stretching of the folded layer will be

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[1] This says nothing more than that the limbs of a competent layer will more likely be sheared across than will the arches of the folds.

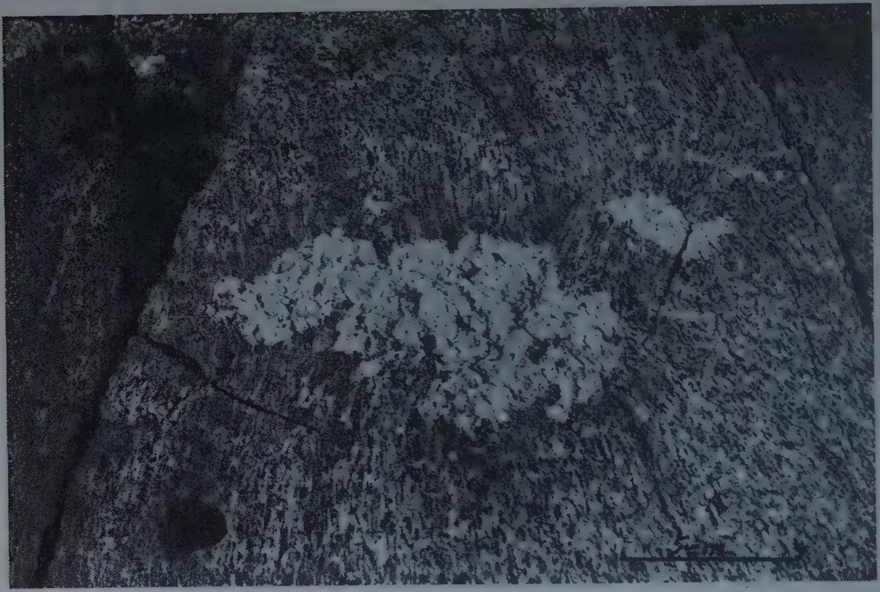

Fig. 10. Pegmatites of type 3. Note change of strike of amphibolite and basified zones adjacent to the pegmatites. Larger pegmatite is 12 cm long. Photo at area A by O. Christie.

greatest at the apex of the fold and on flanks at the points farthest removed from the apices; due to the plastic flow of the quartzite it will be greatest at the points farthest removed from the apices of the folds. These shear and tension directions are shown in fig. 12.

The direction of the preferred orientation of the long axes of the pegmatites at area A corresponds to these shear and tension directions as follows. Of the two oblique shear directions due to compression, one is near parallel to the long axes of the pegmatites. The direction of the tension resulting from the rotational shear is near perpendicular to the long axes of the pegmatites. The direction of the tension due to the shear couple resulting from the movement of the surrounding quartzite layers with respect to the amphibolite is also near perpendicular to the long axes of the pegmatites. The direction of the shear across the limbs of a competent layer approaching isoclinal folding is close to parallel to the regional pressure which itself forms a small angle with the direction of the long axes of the pegmatites, i.e., this shear direction is sub-parallel to the long axes of the pegmatites.

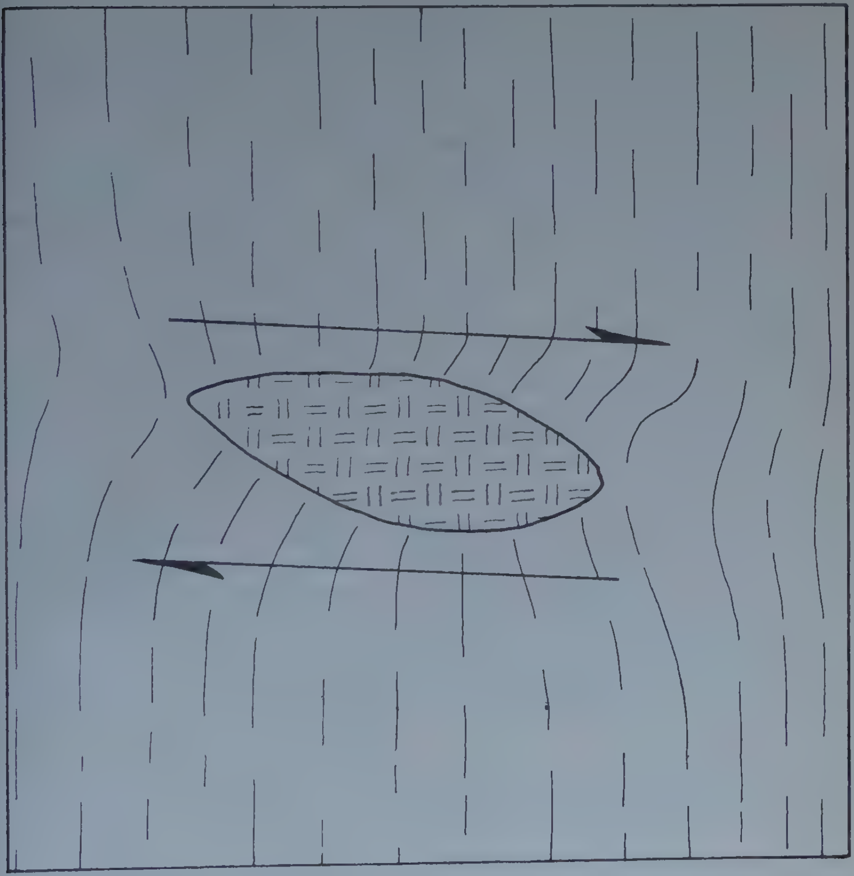

Fig. 11. Simplified and slightly schematized sketch from photo, fig. 10, showing the deduced shear couple parallel to the regional pressure. The gneissic structure indicates this shear plus some plastic deformation of the amphibolite.

Tension in the limbs of the fold due to stretching and due to the plastic flow of the quartzite will be approximately parallel to the strike of the amphibolite and therefore not far from perpendicular to the long axes of the cross-cutting pegmatites. There were therefore forces active tending to shear the amphibolite about parallel to the direction of the long axes of the pegmatites and tending to apply tension about perpendicular to this direction. This means that in the zones now occupied by the pegmatites the original minerals were subjected to a stress due to the shear and that these zones were low

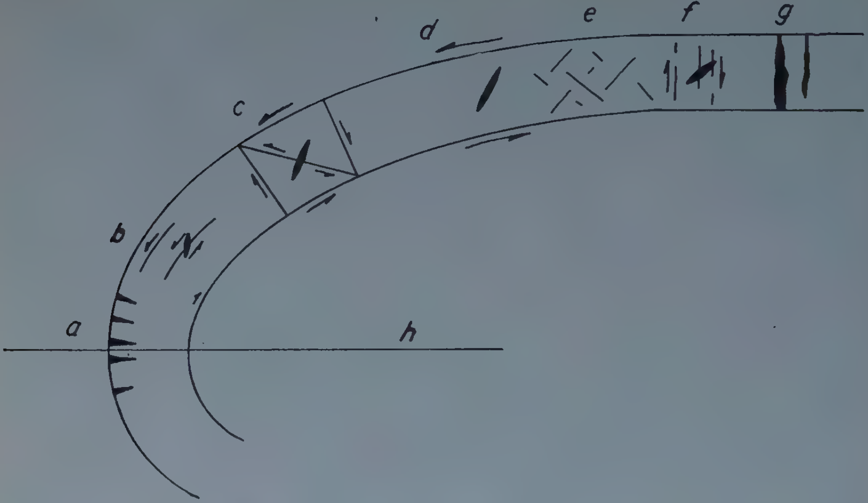

Fig. 12. Shear and tension directions summarized. The arrows show the relative directions of movement. Tension direction is across the black wedges, ellipses, and irregular cracks. a, tension joints; b, concentric shear joints; c, rotational shear; d, shear couple due to relative movement of surrounding layers; e, oblique shear due to compression; f, shear across the competent layer due to regional pressure; g, tension joints due to stretching and to plastic flow of surrounding layers; h, axial plane. Mostly after DE SITTER (7, fig. 72, p. 100).

pressure volumes due to the tension perpendicular to the long boundaries of the zones.

It is well known that crystals which are subjected to a shear-stress have greater chemical activity and tendency to decompose (and crystallize again in more favourable locations or orientations, i.e., recrystallize) than those which are not strained (BARTH, 3, p. 297, 304, MAC DONALD, 11, and RAMBERG, 12, p. 109). In low pressure zones it is also clear that minerals of large mol volume will be formed in preference to minerals of small mol volume, i.e., the relative stabilities of the minerals will be such that those of large mol volume will be favored (see e.g. the discussion in REITAN, 16, p. 302—303).

The pegmatites under consideration are small tabular or pillow shaped bodies. They are completely ductless. Therefore, for the reasons stated by REITAN (15, p. 228—232), magmatic and hydrothermal modes of origin for these pegmatites are excluded as quite unsatisfactory (see also SVERDRUP, 17).

The pegmatites are parallel to directions of shear and the directions of low pressure zones to be expected. It is therefore proposed that these pegmatites formed by a process of metamorphic differentiation which was initiated and controlled by the structural conditions which prevailed. In the zones in which the pegmatites grew the activity of the original minerals was increased by the shear-stress and recrystallization began. Because these same zones were at the same time low pressure zones, the stability of minerals of large mol volume was relatively increased. Concentration gradients were thereby initiated such that material from the diffuse phase in the adjacent rock migrated to the low pressure zones and there crystallized, this material being the constituents of the relatively most stable minerals in the low pressure volumes, i.e., primarily plagioclase and quartz. Of course the composition of the minerals which crystallized in the low pressure volume was dependent on the composition of the environment. K-feldspar, for example, could not grow in the pegmatites, even though the mol volume of K-feldspar is favourably large, because not enough K could be provided to the phase from which the minerals of the pegmatites crystallized by the K-poor amphibolite. As the diffuse phase in the adjacent amphibolite was depleted, material moving down the concentration gradient towards the volume in which the pegmatite grew, minerals in the amphibolite (namely primarily plagioclase and quartz) simultaneously decomposed in order to maintain equilibrium between the solid and the diffuse phases. This process results in a leaching of the adjacent amphibolite of the minerals enriched in the pegmatite, i.e., the light colored minerals plagioclase and quartz. Such leached, dark zones adjacent to the pegmatites can be seen in figs. 1, 9, 10, 13, 14, and 15. These zones were carefully examined in the field and are truly basified zones, not merely darker in appearance due to change of grain size; hornblende is enriched in these zones.

The plagioclase of the pegmatites is less basic than that of the surrounding amphibolite. Such is also the case for other small pegmatites in the Kongsberg—Bamble formation which the author has examined (REITAN, 15, 16) and, as was the case for these other pegmatites, this is most probably due to the diffusion rate of Na being more rapid than that of Ca. Because the pegmatites on Rytterholmen have been very locally derived, the differentiation of Na and Ca has not been very large and the difference between the plagioclase in the

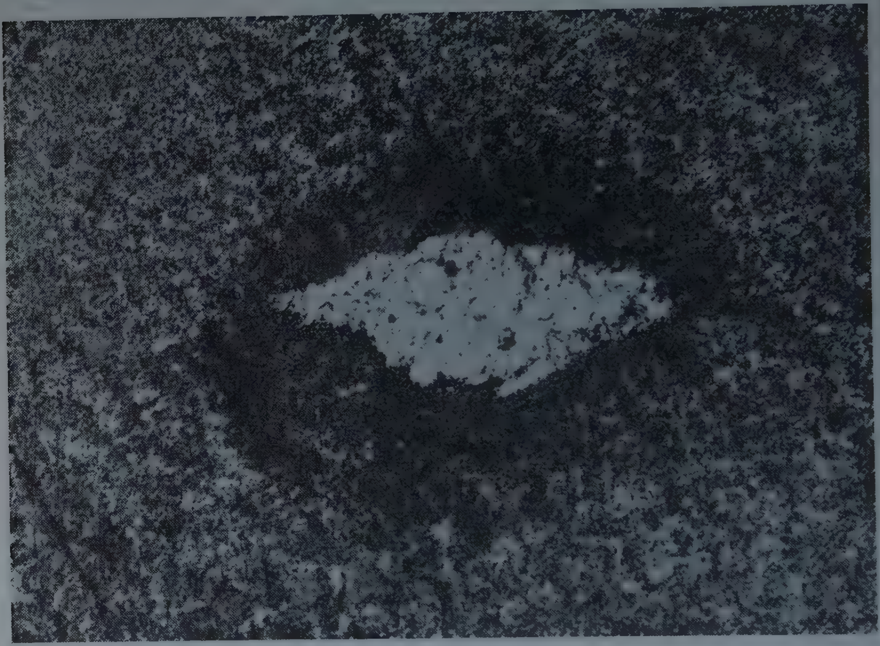

Fig. 13. Unusually strongly marked basified zone around a 20 cm long pegmatite in the amphibolite at area C. Photo by P. Reitan.

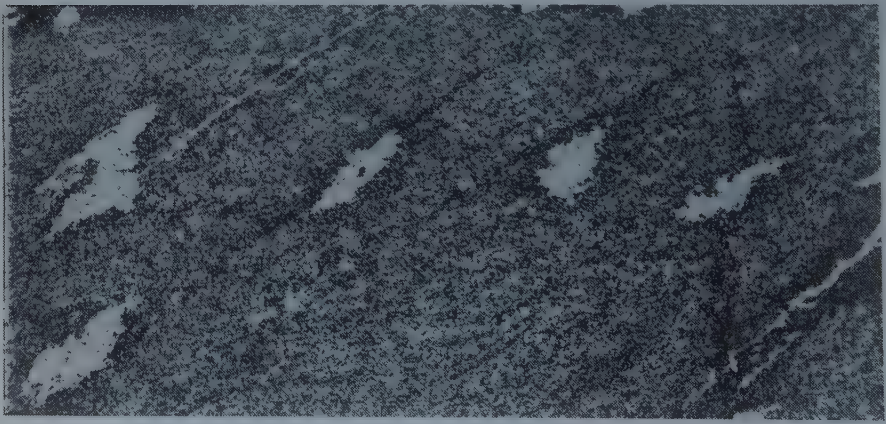

Fig. 14. Basified zones around pegmatites nearly parallel to the strike of the surrounding amphibolite at area E. The pegmatite with the central black spot is 13 cm long. Photo by P. Reitan.

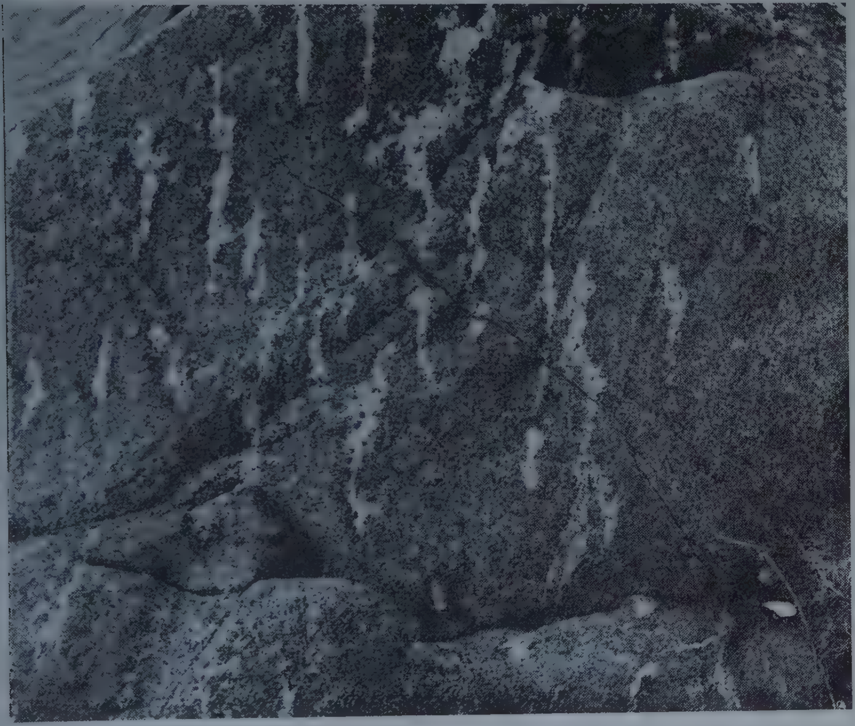

Fig. 15. Elongate pegmatites at area B. The photo shows about 3 m breadth at the bottom. The exposure is almost vertical; the pegmatites have the same shape on horizontal exposures. Photo by P. Reitan.

amphibolite (An_{42-44}) and in the pegmatites (An_{38-40}) is therefore not very large.

Areas B, C, and D are in structural positions completely comparable to that of A, the only differences being that area B is slightly closer to the apex of the fold and areas C and D are slightly farther away. This leads to small differences in the relative magnitude of the processes which tend to cause shear and tension.

The two oblique shear directions due to compression are probably best developed at distances farthest from the apices of the folds. Rotational shear is probably best developed at some point between the apex of the fold and the mid-point of the flank. The intensity of the shear couple due to movement along the boundaries between the amphibolite and the quartzite will increase with distance from the apices of the folds. The shear across the flanks of the amphibolite layer due directly to the regional pressure on the folded competent layer surrounded by relatively plastic material is greatest about midway between the apices of the folds, assuming symmetrical folds in the layer. Tension due to stretching of the folded layer will be greatest at the apex of the fold and on the flanks at points farthest from the apices; due to plastic flow of the quartzite it will be greatest at points farthest removed from the apices of the folds. These factors, along with such things as the local composition of the amphibolite and inhomogeneities in the amphibolite, will combine to determine the number, size, shape, and attitude of the pegmatites.

Area E is in a somewhat different structural position, in that it is so close to the axis of the fold. Here the two oblique shear directions due to compression are minimal. Rotational shear is, though not absent, not intense. The shear couple due to movement along the boundaries of the band approaches its minimum. Shear across the amphibolite due directly to the regional pressure will not be very effective in the arch of a fold in a competent layer surrounded by relatively plastic material. The tension due to stretching will be great at the apex of the fold, will be perpendicular to the tangent to the arc of curvature, and will diminish with distance from the apex. However, in the axial region of the fold another form of shear will play a role of importance, namely concentric shear joints. This, combined with the tension perpendicular to the tangent to the arc of curvature of the fold, will yield zones nearly parallel to the strike in which there exist both the shear-stress on the mineral grains necessary to start recrystallization and a component of tension perpendicular to the direction of the shearing-stress. The result is that the pegmatites in this area will be oriented with their long axes more nearly parallel to the strike of the amphibolite than was the case along the flanks of the fold. Such was also observed to be the case (cf. figs. 6 and 14).

The lengths and the widths of the pegmatites in four of the areas were measured and the ratio of the length to the width calculated. The results of these measurements and calculations are:

Area	l/w max.	l/w min.	l/w average
A	10.0	1.4	4.3
B	14.0	3.7	7.6
C	6.3	1.7	3.4
E	6.7	1.5	3.0

After examining the other data which are available the best correlation seems to be with the amount of light minerals present or, inversely, with the sum of the dark minerals in the amphibolite. Otherwise stated the correlation is inverse to the basicity of the amphibolite. As it is well established that basic rocks are less plastic than more acidic rocks during metamorphism (see e.g. HOFSETH, 9, and RAMBERG, 13 and 14) this is then a positive correlation with the plasticity of the amphibolite. This is of course reasonable. The growth of the pegmatites in the amphibolite is a response to the deformation and will tend to relieve the stress which is their cause. A brittle host rock to the pegmatites will be relatively little capable of plastic deformation, such that the relief of the stress (in this case elongation by plastic flow in the direction of the tension) will tend to be concentrated at the sites of the pegmatites. They will grow relatively more in the direction of tension than their counterparts in a more plastic rock. The pegmatites in the more basic and therefore less plastic host rocks become wider in comparison to their length than those in the more plastic rocks, as the direction of the tension is perpendicular to the length of the pegmatite bodies; the average ratio l/w decreases as the basicity of the amphibolite increases. The observations show that the amphibolite is most basic at areas C and E and the ratio l/w is lowest in these areas, although inverted with respect to the basicity of the amphibolite. However, the average at area E is based on only 15 measurements and is probably the least reliable average. Area A is less basic and has a higher ratio l/w, and area B, at which the least basic amphibolite was found, has the highest ratio l/w.

The process which has been described here to explain the mode of origin of the pegmatites is one which is believed to have taken place during the dynamo-thermal metamorphism of the area, a reaction which was initiated by and which relieved the stress which was created by the deformation. This process is also dependent on elevated temperature; at low temperature the reactions would be too sluggish and relief of the stress would have to occur by some other means. Specifically, in this case it is not believed that there has occurred rupture of the amphibolite at the sites of the pegmatites. The zones were subjected to a shear-stress, but the increased activity induced by the stress in these zones was enough, with such high temperature, to allow the stress to be relieved by a process of recrystallization and metamorphic differentiation before rupture occurred: the growth of new crystals at the sites of the pegmatites kept up completely with the extension of the rock due to tension perpendicular to the shear-stress. To some extent the stress was also relieved by plastic deformation (through recrystallization) of the amphibolite.

The petrographic descriptions of the amphibolite indicate that this area has been subjected to only very insignificant retrograde metamorphism. It is therefore believed that, as there are no visible sheared zones associated with these pegmatites, zones of mechanically ruptured crystals did not exist at the time when the pegmatites were formed. Metamorphic recrystallization relieved the stress which could cause rupture before rupture occurred.

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PEGMATITE VEINS AND THE SURROUNDING ROCKS

IV. Genesis of a discordant pegmatite vein, St. Hansholmen, Risør, Norway

BY

PAUL REITAN

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Abstract: A detailed petrographic study was made of the banded gneiss surrounding a pink, microcline rich, discordant pegmatite vein crossing the western end of the island St. Hansholmen near Risør, southern Norway. The banded gneiss is part of the pre-Cambrian Kongsberg—Bamble formation and consists of alternating, quite regular bands of amphibolite and granodioritic to quartz dioritic gneiss.

Modal analyses of five series' of samples of the gneiss and one of the amphibolite reveal a K metasomatism of the gneiss adjacent to the pegmatite vein, the intensity of which diminishes with distance from the vein. This is rendered even clearer by the calculated cation percent compositions of the samples. The amphibolite has been very little if at all affected by the emplacement of the vein.

Volume change calculations indicate that the material which constitutes the vein entered into a crack across the banded gneiss. It is concluded that the material which filled the crack was derived from the nearby granulite facies rocks.

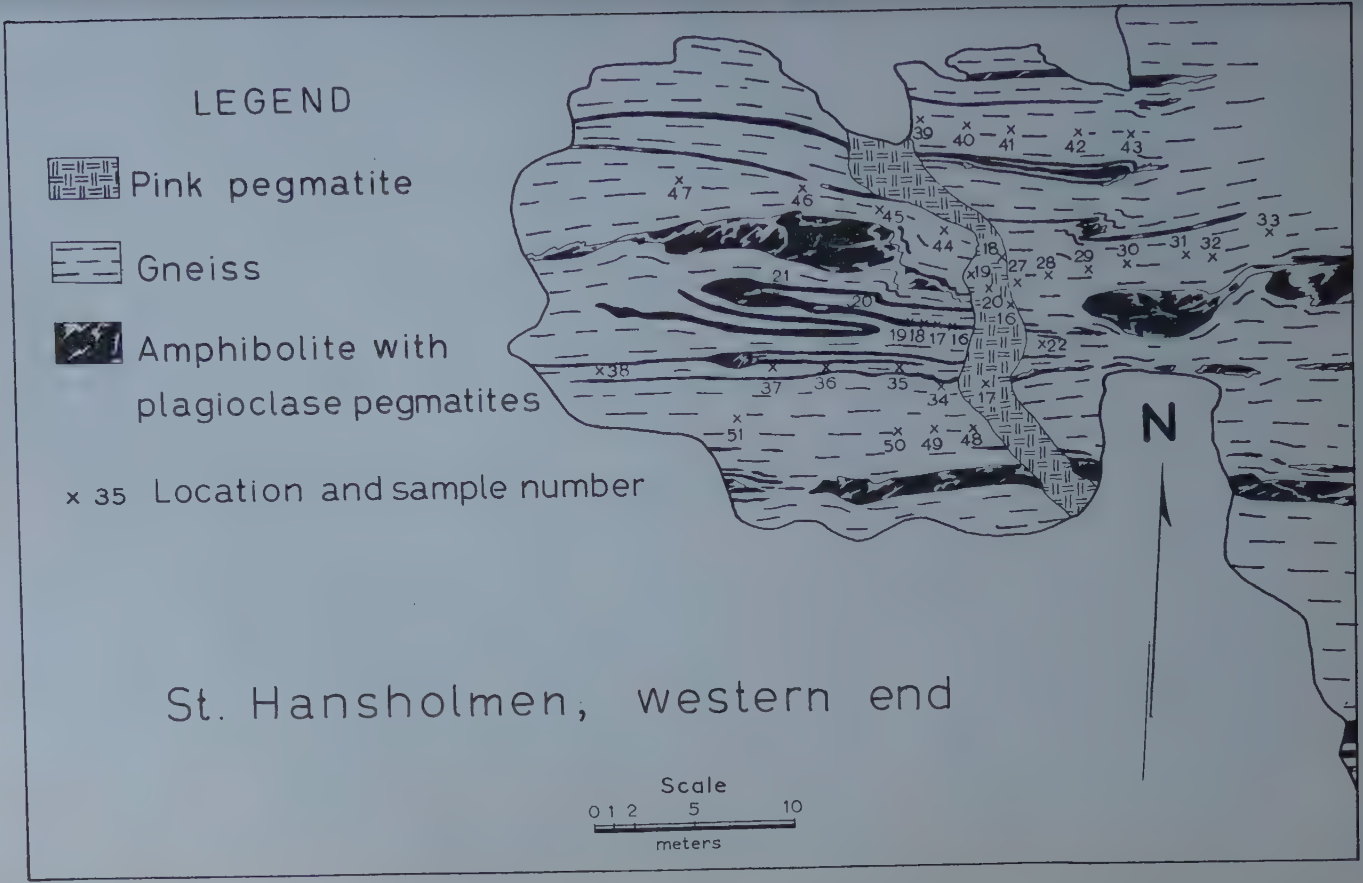
Temperature determinations using the feldspar thermometer, both of the pegmatite and the gneiss, indicated that the pegmatite formed at a higher temperature than the surrounding gneiss (ca. 570° C vs. 460° C), although lack of equilibrium between co-existing feldspars in the gneiss near the vein makes it impossible to determine temperature differences in the gneiss as related to distance from the vein, if such existed.

The nature of the pegmatite forming fluid is considered. Four possibilities are discussed: 1) magma; 2) hydrothermal solution; 3) fluidized gas-solid system; and 4) disperse phase of relatively high energy particles. Possibility 2) is discarded because of the large quantity of water which would be required and because 10 km^3 of rock would have had to have been «dehydrated» to yield solution sufficient to form a vein 500 m × 2 m × 500 m. Possibly 3) is not advocated because none of the characteristic features of veins formed by such a system have been found. None of the observations which were made are found to constitute either decisive proof for or evidence fatal to possibilities 1) and 4); no conclusion can be drawn except that the vein formed from a fluid (magma or disperse phase) «injected» into a crack in the banded gneiss.

Introduction

In this paper the pink, microcline rich pegmatite crossing the western end of the island St. Hansholmen, which lies about 1 km from the town Risør on the south coast of Norway, is described; the changes in and the relations of the vein to the surrounding banded gneiss are discussed. This pegmatite vein was mentioned in the first paper of this series (Reitan, 12), although discussion of the genesis of the vein was deferred until such time as detailed studies of the pegmatite and the immediately adjacent rocks had been made. The petrography and structure of the banded gneiss on St. Hansholmen and the genesis of the very small, white, plagioclase pegmatites within and between the boudins of the amphibolite bands have been discussed previously (Reitan, 12). The most detailed geological map of the Søndeled district (including Risør) is by J. A. W. Bugge (3, fig. 3, p. 13). Other literature pertinent to the geology of the Kongsberg—Bamble formation has been listed previously in this series (Reitan, 12).

As is the case on other parts of St. Hansholmen, the dips in the area near the pink pegmatite crossing the western tip of the island are essentially vertical, as are the dips of the contacts between the pegmatite vein and the banded gneiss. The distribution of the rock types within the area under consideration and the locations at which samples were taken and upon which the petrographic and chemical data are based are shown in the map, fig. 1.



St. Hansholmen, western end

Petrography

Petrographic descriptions of the gneissic and amphibolitic bands of the banded gneiss and of the small plagioclase pegmatites have been given earlier (REITAN, 12); there the characteristic features and variations of these rocks are discussed. The banded gneiss at the western end of St. Hansholmen is essentially similar to that of the rest of the island; the petrographic descriptions here will be primarily for the purpose of elucidating changes in the gneiss as related to distance from the pegmatite vein.

PEGMATITE

The cross-cutting, pink pegmatite vein at the western end of the island varies between 1 and 3 meters in width. Microcline occurs in crystals up to 10 cm in largest dimension; it contains very fine perthite lamellae. Biotite occurs in plates up to ca. 4 cm. Plagioclase occurs in unzoned grains up to 2 cm. Quartz occurs in small grains graphically intergrown in microcline. Hornblende has been seen in a few grains up to 2 cm long. Microcline is the overwhelmingly dominant mineral in the pegmatite vein.

Changes in the composition of the feldspars in the pegmatite with distance from the nearest boundary have been observed. Microcline from three samples of the pegmatite has been analyzed for K_2O and Na_2O : from sample ÖP19 (ca. 10 cm from the boundary) it contains 80.7 Or and 19.3 Ab (molecular proportions); from ÖP18 (ca. 40 cm from the boundary) it contains 79.8 Or and 20.2 Ab; from ÖP17 (ca. 120 cm from the boundary) it contains 78.0 Or and 22.0 Ab. The plagioclase from samples ÖP16, ÖP18, and ÖP19 (all within 50 cm of the vein boundary) contains 31 An and 69 Ab while the plagioclase from samples ÖP17 and ÖP20 (both about at the center of the vein) contains 27 An and 73 Ab. In other words, both feldspars become more Na rich towards the center of the vein.

The triclinicity of the microcline from three samples from the pegmatite was determined (ÖP17—19). It was found to be very high (see table 2), indicating a high degree of Si—Al order in the microcline lattice. No differences other than those attributable to experimental error could be detected.

No zoning, either in terms of grain size, textural relations, or mineralogy could be observed in the pegmatite, except just at the vein boundaries where the pegmatite becomes quite abruptly more fine-grained, although there is no line of demarkation. The contacts with the gneiss are gradational over a few centimeters, but against the amphibolite bands they are absolutely sharp.

BANDED GNEISS

The banded gneiss consists of dark bands of amphibolite and light bands of gneiss of granodioritic to quartz dioritic composition. The light bands are generally thicker than the dark bands, though both are quite variable. Within the dark bands there are numerous, small, white, plagioclase pegmatites. White, plagioclase pegmatites are also frequently found between boundins of the amphibolite bands (cf. map and description, REITAN, 12).

The gneissic bands near the pegmatite vein contain the following minerals: quartz, always; plagioclase, always; microcline, almost always though in highly variable amounts; biotite (partly altered to pennine), always; zircon, always; opaques (mostly magnetite), almost always; garnet, sometimes; apatite, sometimes; muscovite, rarely; and epidote, rarely. The amphibolite contains hornblende, always; plagioclase, always; biotite (partly altered to pennine), always; opaques (mostly magnetite), always; apatite, always; and quartz, almost always.

Five series' of samples from gneissic bands and one from an amplibolite band were taken. Modal analyses of these samples were made in order to determine the quantitative mineralogical variations in the rocks surrounding the pegmatite vein as a function of distance from the vein. The results of these modal analyses are presented in table 1 and graphically in figs. 2 to 7. For the locations of these samples see the map, fig. 1.

Certain general trends are apparent and some of the deviations from regularity can be accounted for.

In all of the series' from the gneiss the amount of microcline is greatest nearest the pegmatite vein. The amount of quartz seems to bear no consistent relationship to distance from the vein. The same can be said of biotite. The amount of plagioclase varies essentially

Table 1 Modal analyses of thin-sections

Sample no.	Plagioclase				Quartz	Micro-cline	Biotite (+ chlo-rite)	Muscovite	Horn-blende	Opaques	Zircon	Garnet	Apatite	Epidote
	An[1]	Total	Albite	Anorthite										
ÖG 22	20	19.4	15.5	3.9	43.2	28.0	9.2			—	—			—
27	20	15.2	12.2	3.0	43.1	33.9	7.7			—	—			
28	27	44.4	32.4	12.0	47.2	1.7	5.1	1.5			0.1			
29	30	43.2	30.2	13.0	42.0	11.6	2.8	0.3			0.1			
30	30	33.9	23.7	10.2	42.2	17.8	6.0			0.1	—			
31	24	35.0	26.6	8.4	40.6	18.8	5.6			—	—			
32	29	50.3	35.7	14.6	38.7	5.0	5.9				—			
33	30	36.7	25.7	11.0	47.1	8.6	7.5			—	0.1			
34	20	30.6	23.9	6.7	45.7	18.1	5.5			—	—		—	
35	29	47.3	33.8	13.5	43.9	2.6	5.5			0.4	0.1	0.2		
36	32	49.1	33.4	15.7	45.5	0.4	4.5			0.5	—			—
37	20	46.8	37.4	9.4	45.3		6.8			0.7	0.1	0.3	0.1	
38	22	51.3	40.0	11.3	42.7	0.7	4.8			0.5	—		—	
39	25	20.2	15.2	5.0	42.0	30.0	7.0			—	0.2	0.5		
40	28	28.3	20.4	7.9	43.7	17.8	9.7			—	0.2	0.2	0.1	
41	29	30.8	21.9	8.9	43.7	18.9	6.0			0.2	0.2	0.3		
42	30	31.0	21.7	9.3	44.1	19.7	4.8			0.1	—	0.3		
43	29	36.4	25.9	10.5	40.9	13.2	8.3	0.2		0.1	0.1	0.8		
44	20	24.0	19.2	4.8	39.5	29.6	6.8			—	—		0.1	
45	28	38.3	27.6	10.7	42.5	13.8	5.2			—	0.2			
46	28	49.5	35.9	13.6	44.8	0.2	4.9			0.4	0.2			
47	29	32.0	22.6	9.4	48.0	14.8	4.6			0.5	0.1			
48	20	24.4	19.5	4.9	36.7	32.4	6.3			0.2	—			
49	28	43.6	31.4	12.2	43.9	6.2	6.2			—	0.1			
50	29	48.5	34.2	14.3	38.7	3.9	8.0			0.4	0.1	0.4		
51	31	48.7	33.4	15.3	42.7	2.7	4.2			1.6	0.1			—
ÖA 16	32	36.4	24.8	11.6	0.5		3.3		58.3	1.2			0.3	
17	32	32.2	22.2	10.0	0.9		4.5		61.0	1.1			0.3	
18	32	32.3	22.3	10.0	0.2		1.5		64.4	1.3			0.3	
19	32	39.0	26.8	12.2	0.3		1.4		57.6	1.6			0.1	
20	32	37.4	25.7	11.7	0.6		0.3		59.7	1.8			0.2	
21	30	33.0	23.1	9.9			0.1		65.6	1.0			0.3	

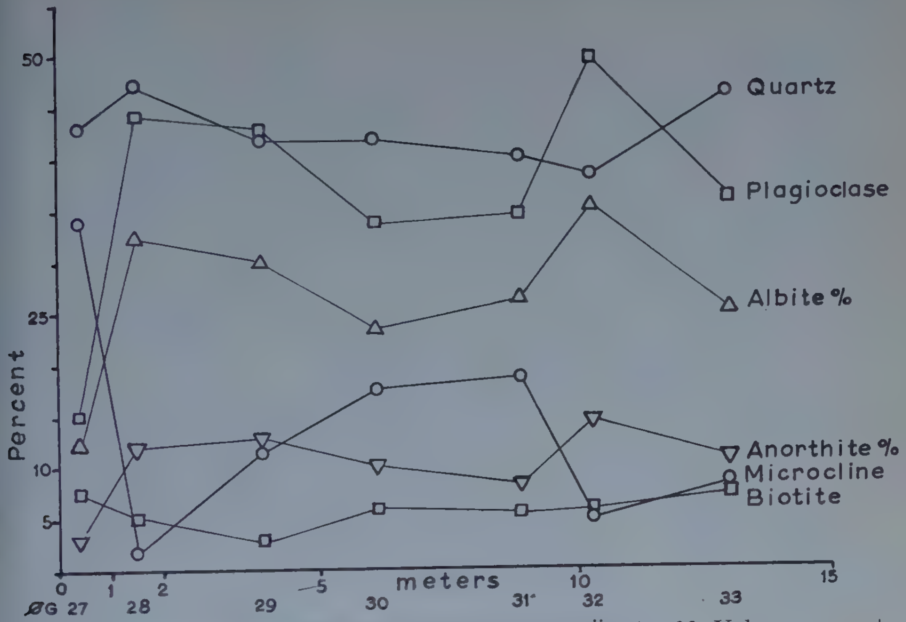


Fig. 2. Modal analyses of series of gneiss samples ÖG27—33. Volume percent vertically; distance from pegmatite horizontally.

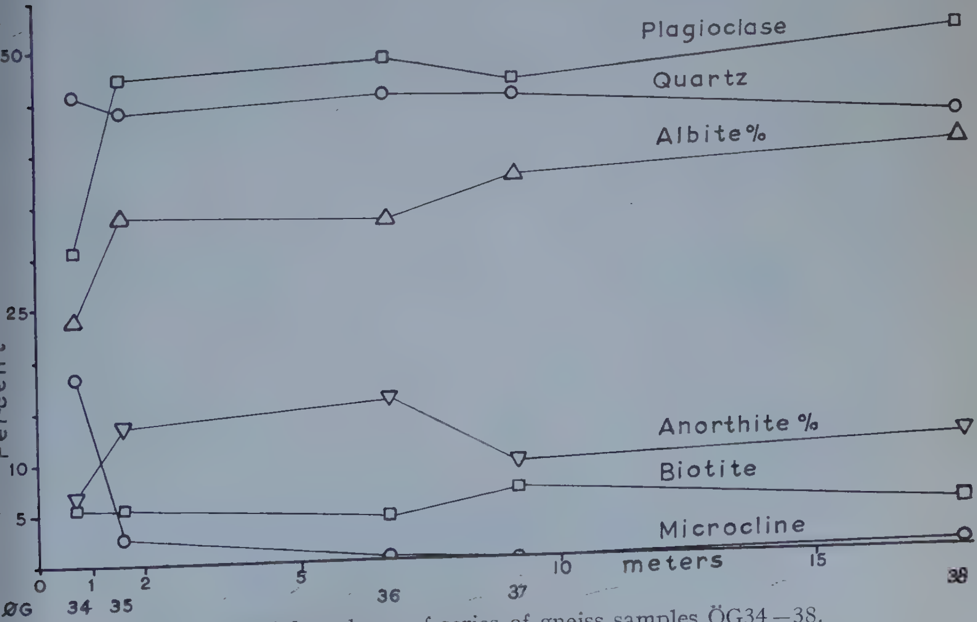


Fig. 3. Modal analyses of series of gneiss samples ÖG34—38.

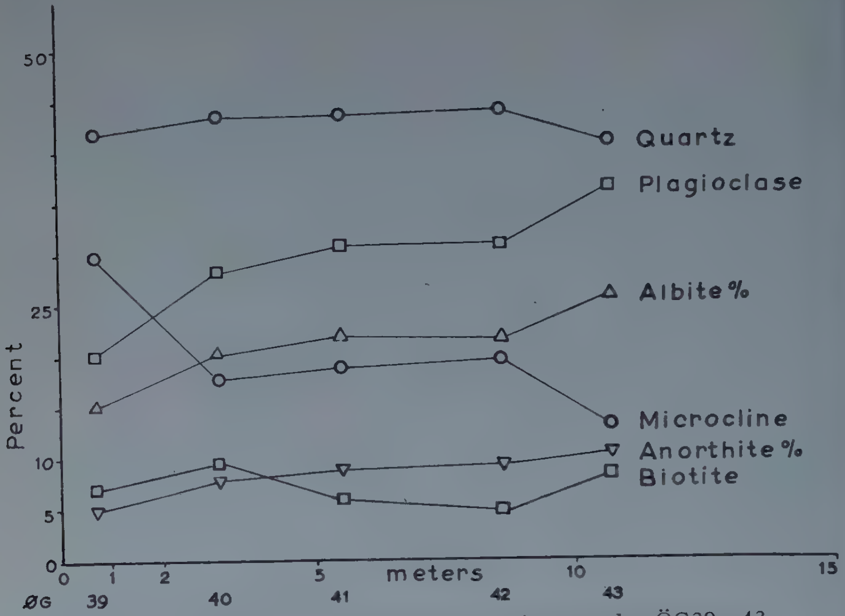


Fig. 4. Modal analyses of series of gneiss samples ÖG39—43.

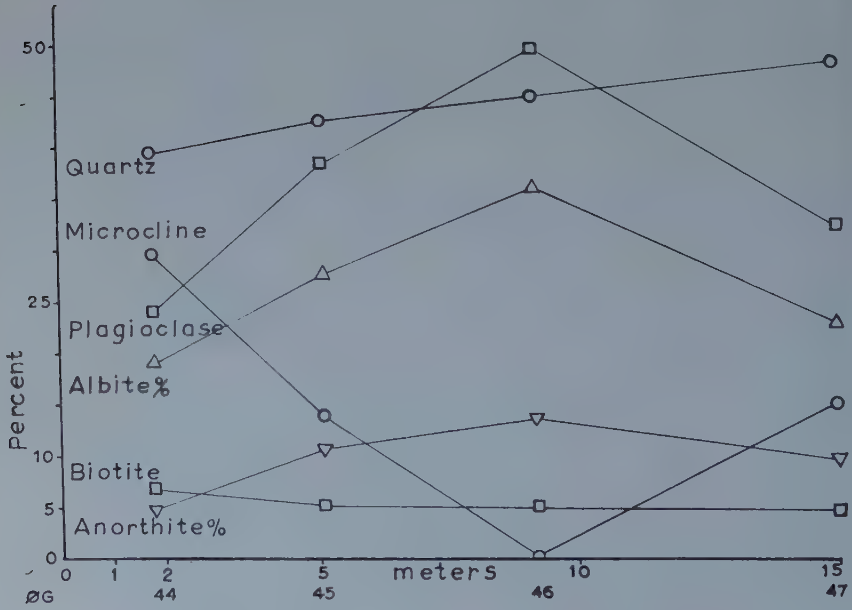


Fig. 5. Modal analyses of series of gneiss samples ÖG44—47.

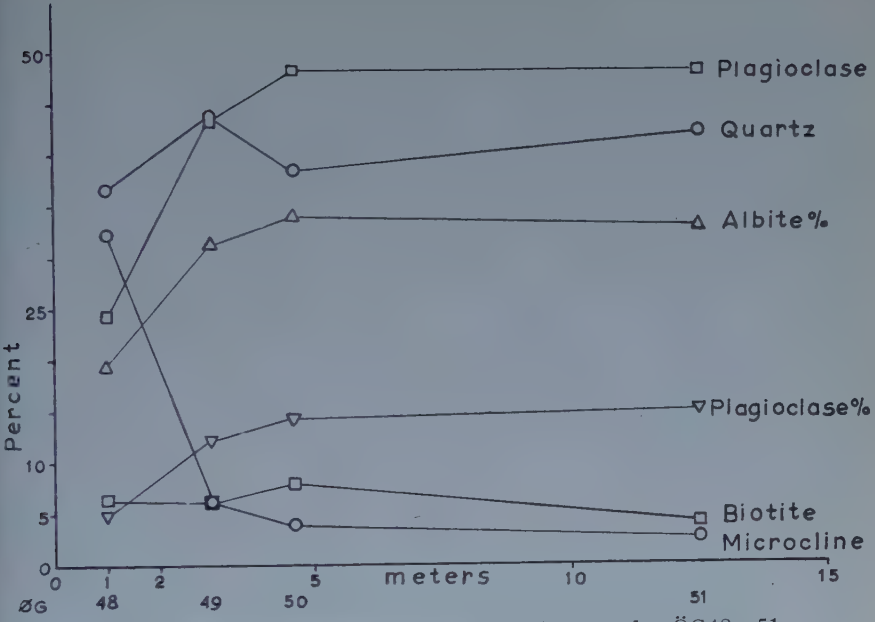


Fig. 6. Modal analyses of series of gneiss samples ÖG48—51.

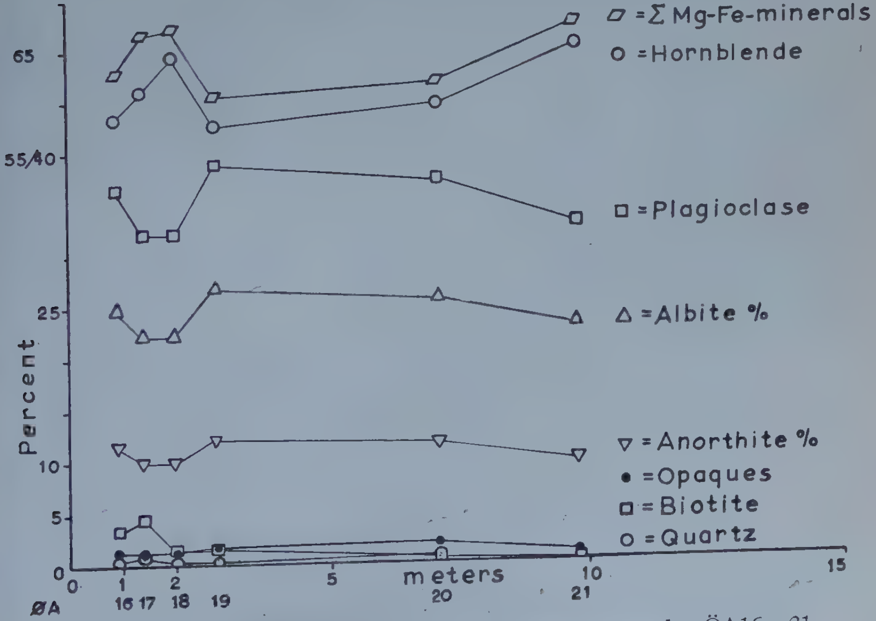


Fig. 7. Modal analyses of series of amphibolite samples ÖA16—21.

complimentarily to that of microcline and is always least nearest the pegmatite vein.

The amount of microcline in the gneissic bands is less near the amphibolite bands than in the middle of the gneissic bands (REITAN, 12). For this reason the amount of microcline is very low in the series of samples ÖG35—38, all of which were taken very close to the boundary of an amphibolite band. Sample ÖG34 was also taken very close to the boundary of an amphibolite band but within 1 m of the pegmatite vein, and although the amount of microcline is appreciable it is only a little more than half of that found in the other samples from within 1 m of the pegmatite vein. The amount of microcline in sample ÖG46 is also low; it was taken close to an amphibolite band. The reason for the amount of microcline being as low as it is in sample ÖG28 is unknown.

The only variation in the series of samples from an amphibolite band to which any degree of significance could possibly be attached is the slight increase of biotite near the pegmatite vein. The variations of the other minerals with distance from the vein are inconsistent and the maximum deviations of the major minerals from the average amount present are 10 % or less of the amount recorded.[1]

Certain textural features of the gneiss are worthy of mention. The grain size of the gneiss is generally medium- to medium-fine-grained, although quartz sometimes occurs in quite large grains. The grain size of microcline is highly variable; close to the vein it always occurs both as large grains and as interstitial grains and stringers, then usually between quartz and plagioclase. In the sections within 0.7 m of the vein microcline appears to be in part replacing plagioclase, as occasional microcline grains can be seen to include small plagioclase grains which have the same optical orientation and may have the same orientation as an adjacent plagioclase grain; they appear to be unreplaced remnants of a formerly larger plagioclase grain. With increased distance from the vein the microcline occurs as interstitial grains and stringers which follow the grain boundaries of the larger grains and seem to occur preferentially along the boundaries between quartz and plagioclase. When the total amount of microcline exceeds about 13 percent there occur irregularly shaped grains (not interstitial stringers) as well as tiny interstitial grains and stringers. When the total amount of microcline is below 10 percent it occurs predominantly as interstitial grains and stringers. This would suggest the possibility that in the gneiss bands there is a variable amount of original microcline and a certain amount of secondary microcline which diminishes with distance from the vein. The above observations might be attributable entirely to secondarily grown microcline which, when in sufficient quantity, also forms non-interstitial grains; however, observations from the gneiss from other parts of St. Hansholmen (REITAN, 12) would tend to support the suggestion that the microcline in the gneiss near the pink pegmatite vein belongs to two generations. The older is variable in amount, varying both from band to band and within each band as a function of distance from the amphibolite bands; the younger in part replaces plagioclase near the vein and otherwise occurs as irregular grains and interstitial grains and stringers, the total quantity of microcline diminishing as the distance from the veins increases until the secondary microcline present occurs only as tiny interstitial grains and stringers.

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[1] The curves for plagioclase, hornblende, and sum of dark minerals resemble those found in the partly amphibolitized olivine gabbro surrounding three pegmatite veins near Risør (REITAN, 13), but the deviations from the average in the amphibolite band are significantly smaller (10 % as compared to about 30 %) and result in only very small changes in the calculated chemical compositions (compare table 4 and fig. 13 with REITAN, 13, p. 291, 296, 297).

The triclinicity of the microcline from six samples from the gneiss was measured (table 2, ÖG27, 29—33). No differences other than those attributable to experimental error could be measured. In all cases the degree of triclinicity is quite high (though lower than that found in the pegmatite), indicating a high degree of Si—Al order in the lattice.

The degree of sericitization[1] of the plagioclase in the gneiss is not uniform. In general, the sericitization within about 2 m of the vein is intense; further from the vein the sericitization is somewhat variable though always rather slight. The composition of the plagioclase is variable.[2] The samples taken nearest the veins contain relatively albite rich plagioclase (An_{20} except for ÖG39 which is An_{25}). Those

[1] That the alteration product is sericite and not paragonite is indicated by the fact that it (along with microcline, muscovite, and biotite) became yellow after the staining process to which the thin sections were subjected.

[2] Variable from sample to sample. Along plagioclase-microcline contacts there can frequently be seen an extremely thin, unsericitized rim on the plagioclase grain (not always continuous) which has a very slightly different extinction angle and probably slightly lower index. The rims are so thin that no quantitative differences could be determined.

Table 2.

Triclinicity of microcline

Sample no.	Rock type	Distance from contact	Triclinicity △
ÖG 27	Gneiss	0.3 m	0.85
29	»	3.9 m	0.85
30	»	6.1 m	0.85
31	»	8.9 m	0.85
32	»	10.3 m	0.80
33	»	13.0 m	0.80
ÖP 17	Pegmatite	1.2 m	0.95
18	»	0.4 m	0.95
19	»	0.1 m	0.98

more than 2 m from the veins contain more calcic plagioclase which, with three exceptions, lies between An_{28} and An_{32}.

Few changes in the minerals of the amphibolite can be noted. The plagioclase is uniformly An_{32} except for that from sample ÖA21 in which the plagioclase was determined to be An_{30}; i.e. the plagioclase is of essentially uniform composition. Sericitization of the plagioclase is not uniform; in the sample nearest the vein the sericitization is generally strong, in the next sample it is patchy and moderately strong, and in the remaining samples it is strong except for the one most distant from the vein where it is patchy but locally very strong. The sample most distant from the vein is medium-fine-grained, the others fine-grained, and the hornblende in the sample most distant from the vein is slightly lighter in color and not quite so strongly pleochroic as that from the remaining samples. Very thin quartz veinlets are occasionally to be seen in the amphibolite.

Petrochemical calculations

In order to facilitate the comparison of changes within the series' of samples and between series' and in order to elucidate the chemical changes which the observed mineralogical changes imply, the cation percent composition of the samples was calculated on the basis of the modal analyses of the samples.

Table 3

Mineral formulas used in petrochemical calculations.

Quartz	SiO_2
Microcline	$KAlSi_3O_8$
Albite	$NaAlSi_3O_8$
Anorthite	$CaAl_2Si_2O_8$
Biotite[1]	$KAlMg_3Si_3O_{10}(OH)_2$
Hornblende[1]	$Ca_2Mg_4Al_2Si_7O_{22}(OH)_2$
Muscovite	$KAl_3Si_3O_{10}(OH)_2$
Opaques	Fe_3O_4

[1] The formulas used for these minerals are those suggested by Barth (1) for use in calculating the mesonorm, with Fe substituting for Mg and other simplifications.

To do this standard formulas for the minerals involved were used. These are listed in table 3. The amounts of albite and anorthite are dependent on the composition of the plagioclase in the sample as determined optically. The formulas used for biotite and hornblende are those suggested by BARTH (1) for use in calculating the mesonorm from chemical analyses.

All of the formulas are simplified; the distribution of Mg, Fe, and Mn is ignored, Na in microcline is ignored, K in plagioclase is ignored, alkalis in hornblende are ignored, and all formulas are ideal. Accessory minerals are also ignored in the calculations. No claim is made for exact reproduction of the actual chemical composition of the rock. However, it is believed that these calculations make important changes in the chemistry of the rocks evident, much more so than would be the case simply by inspecting the modal analyses, and comparison is considerably facilitated.

The method of calculation is the reverse of a norm calculation, stopping at cation percents. The results of these calculations are given in table 4 and are shown graphically in figs. 8 to 13 where the changes in cation percent are plotted against the distance of the sample from the pegmatite vein; the distance is measured along the strike.

In table 5 the cation percent of the sample closest to the vein in each of the series' is compared with the average percent of each of the cations obtained from the remaining samples of the series.

Table 4

Calculated compositions in cation %

Sample no	Si	Al	Mg Fe	Ca	Na	K
ÖG 22	74.6	11.4	3.4	0.8	3.1	6.7
27	74.8	11.4	3.0	0.6	2.4	7.8
28	75.2	12.9	1.9	2.4	6.5	1.2
29	73.6	13.9	1.1	2.6	6.1	2.7
30	73.6	13.1	2.2	2.0	4.7	4.4
31	73.3	13.1	2.1	1.7	5.3	4.5
32	71.5	14.7	2.1	2.9	7.1	1.7
33	75.2	12.2	2.7	2.2	5.1	2.6
34	75.8	11.7	2.1	1.3	4.8	4.3
35	73.7	13.5	2.1	2.7	6.8	1.2
36	74.0	13.7	1.8	3.1	6.7	0.7
37	74.9	12.2	2.5	1.9	7.6	0.9
38	74.0	13.3	1.8	2.2	8.0	0.7
39	74.3	12.0	2.7	1.0	3.0	7.0
40	73.8	12.1	3.6	1.6	4.1	4.8
41	74.4	12.5	2.3	1.8	4.4	4.6
42	74.7	12.6	1.8	1.9	4.4	4.6
43	72.5	13.2	3.1	2.1	5.3	3.8
44	73.4	12.4	2.6	1.0	3.8	6.8
45	73.7	13.2	2.0	2.1	5.5	3.5
46	74.3	13.3	1.8	2.7	7.3	0.6
47	76.3	12.0	1.7	1.9	4.5	3.6
48	72.2	13.2	2.4	1.0	3.9	7.3
49	73.8	13.2	2.3	2.4	6.3	2.0
50	71.0	14.4	3.0	2.9	6.9	1.8
51	75.3	10.8	3.1	3.1	6.7	1.0
ÖA 16	48.6	17.9	18.0	10.1	5.0	0.4
17	48.6	17.2	19.0	10.1	4.5	0.6
18	48.5	17.3	19.0	10.5	4.5	0.2
19	48.7	18.2	17.4	10.1	5.4	0.2
20	48.8	17.9	17.9	10.3	5.1	—
21	48.6	17.4	18.6	10.7	4.6	

The following differences can be noted for the samples from the gneiss. K is always higher near the vein than the average or the maximum values of the rest of the series. Na and Ca are always lower than the average or the minimum values of the rest of the series. Mg + Fe is in two cases higher than the average and the maximum values

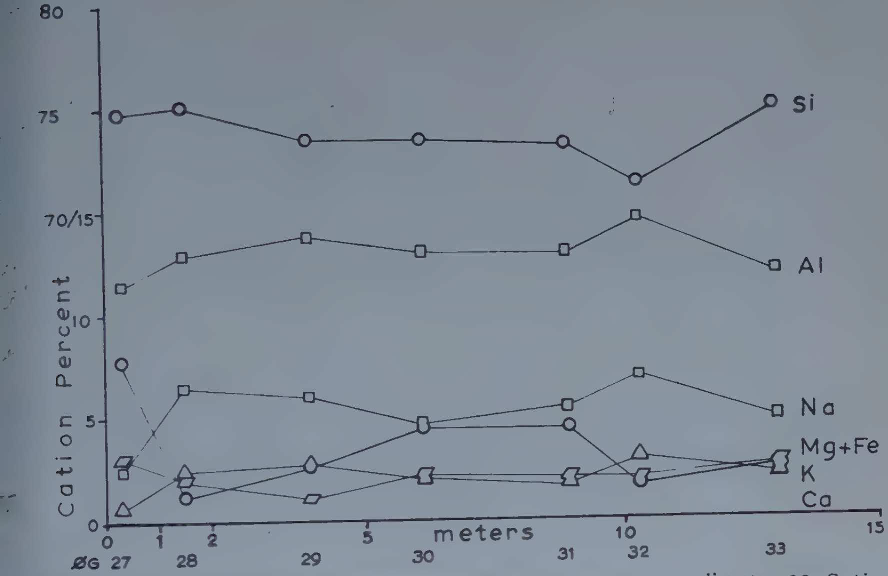

Fig. 8. Calculated cation percent composition of gneiss series ÖG27—33. Cation percent vertically; distance from pegmatite horizontally.

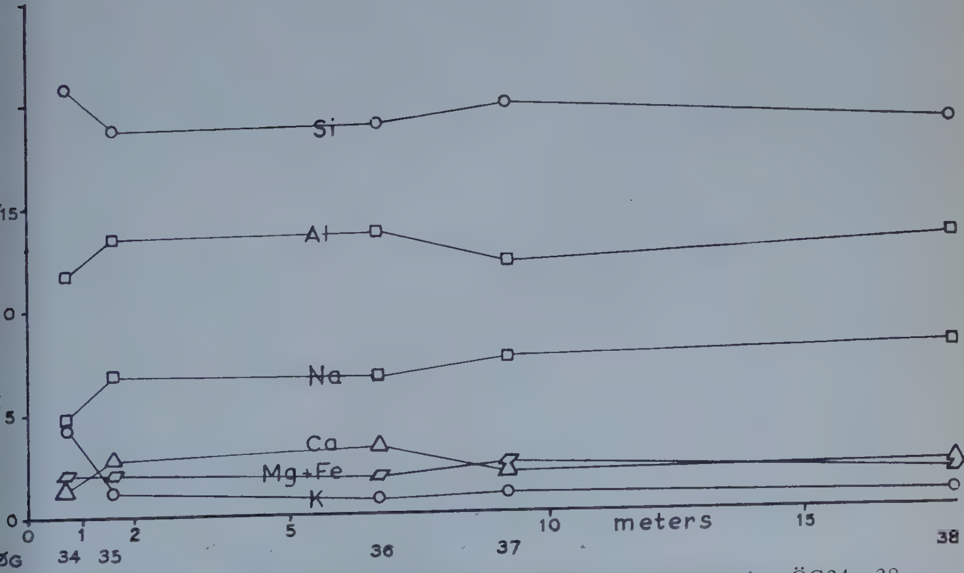

Fig. 9. Calculated cation percent composition of gneiss series ÖG34—38.

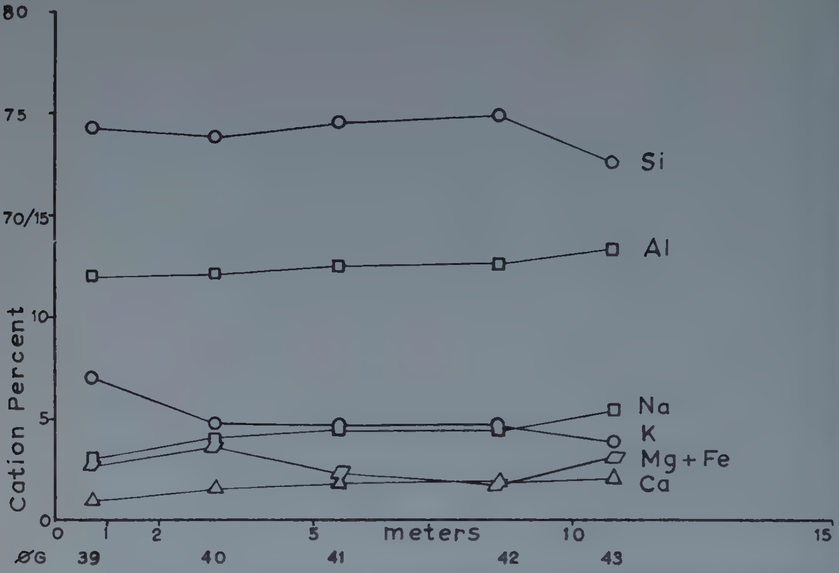

Fig. 10. Calculated cation percent composition of gneiss series ÖG39—43.

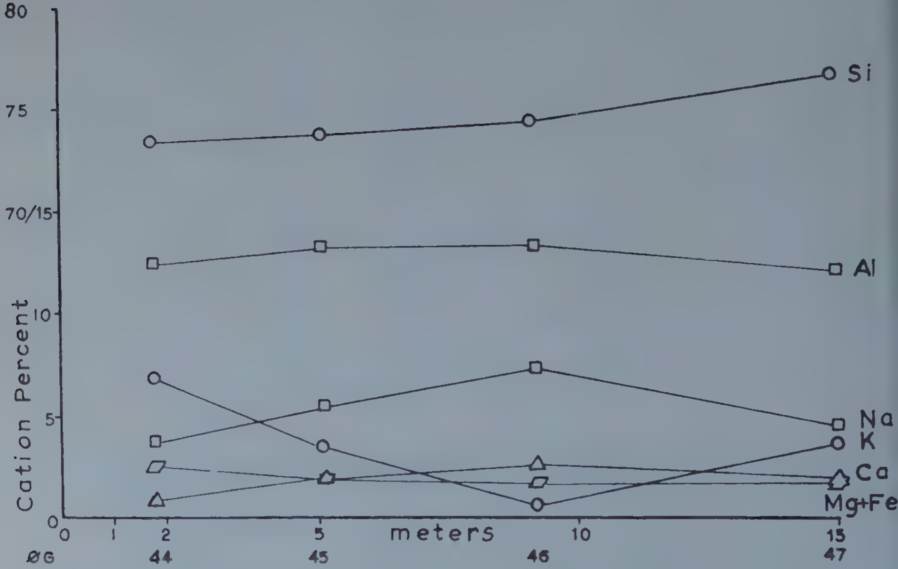

Fig. 11. Calculated cation percent composition of gneiss series ÖG44—47.

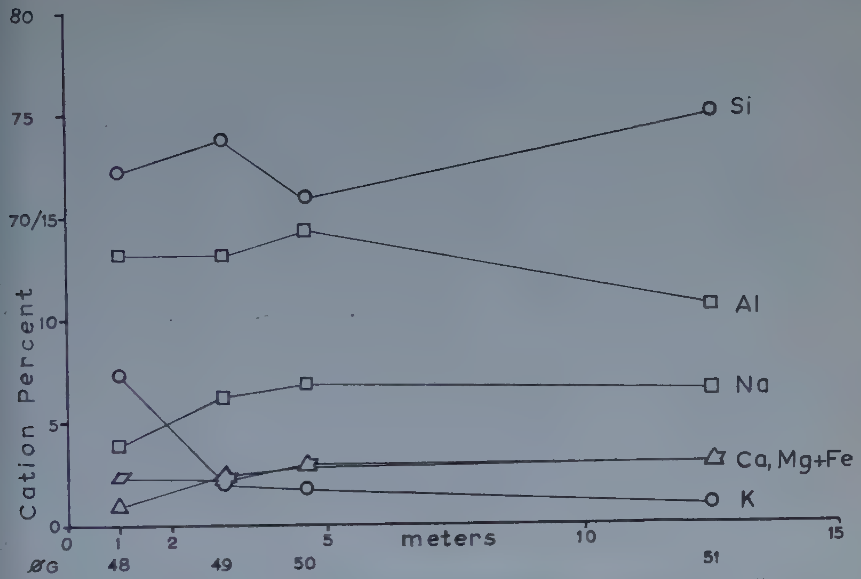

Fig. 12. Calculated cation percent composition of gneiss series ÖG48—51.

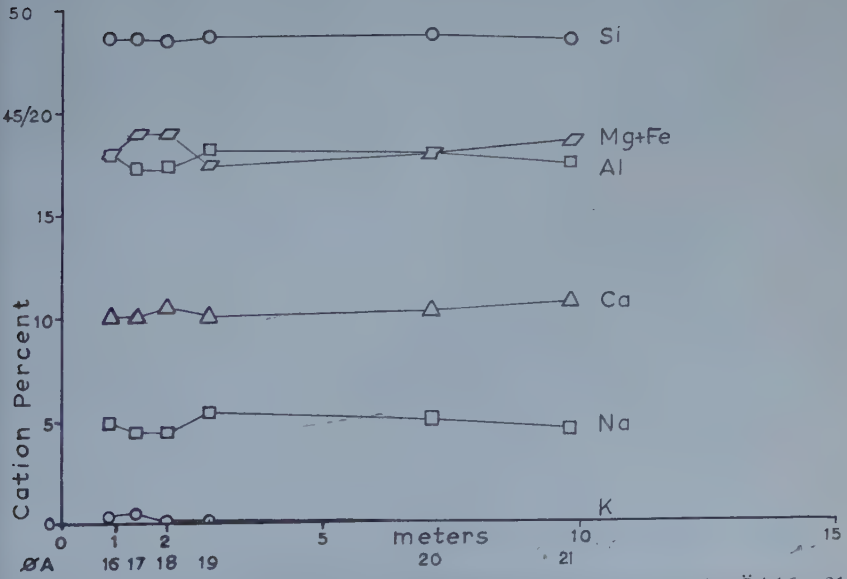

Fig. 13. Calculated cation percent composition of amphibolite series ÖA16—21.

Table 5

Comparison: samples closest pegmatite vs. others.

Sample Series			Si	Al	Mg Fe	Ca	Na	K
ÖG 27—33	28	Max.	75.2	14.7	2.7	2.9	7.1	4.5
	to	Min.	71.5	12.2	1.1	1.7	4.7	1.2
	33	Av.	73.6	13.3	2.0	2.3	5.8	2.8
	ÖG 27		74.8	11.4	3.0	0.6	2.4	7.8
ÖG 34—38	35	Max.	74.9	13.7	2.5	3.1	8.0	1.2
	to	Min.	73.7	12.2	1.8	1.9	6.7	0.7
	38	Av.	74.2	13.2	2.2	2.5	7.3	0.9
	ÖG 34		75.8	11.7	2.1	1.3	4.8	4.3
ÖG 39—43	40	Max.	74.7	13.2	3.6	2.1	5.3	4.8
	to	Min.	72.5	12.1	1.8	1.6	4.1	3.8
	43	Av.	73.8	12.6	2.7	1.8	4.6	4.4
	ÖG 39		74.3	12.0	2.7	1.0	3.0	7.0
ÖG 44—47	45	Max.	76.3	13.3	2.0	2.7	7.3	3.6
	to	Min.	73.7	12.0	1.7	1.9	4.5	0.6
	47	Av.	74.7	12.8	1.8	2.2	5.8	2.6
	ÖG 44		73.4	12.4	2.6	1.0	3.8	6.8
ÖG 48—51	49	Max.	75.3	14.4	3.1	3.1	6.9	2.0
	to	Min.	71.0	10.8	2.3	2.4	6.3	1.0
	51	Av.	73.4	12.6	2.6	2.6	6.6	1.6
	ÖG 48		72.2	13.2	2.4	1.0	3.9	7.3
ÖA 16—21	17	Max.	48.8	18.2	19.0	10.7	5.4	0.6
	to	Min.	48.5	17.2	17.4	10.1	4.5	<0.1
	21	Av.	48.6	17.6	18.4	10.4	4.8	0.2
	ÖA 16		48.6	17.9	18.0	10.1	5.0	0.4

of the rest of the series, but in three cases it is lower than or equal to the average and within the extreme values of the rest of the series. Al is in four cases lower and one case higher than the average, in three cases lower and in two cases within the extreme values of the rest of the series. Si is in three cases higher and in two cases lower than the average, in one case higher, another case lower, and in three cases within the extreme values of the rest of the series.

This indicates that there is a real increase in the percentage of K near the vein and a real decrease in the percentage of Na and Ca.

Table 6

Alkalis in Amphibolite

Sample no.	K_2O Calc. mol. %	K_2O Wt. %	Na_2O Calc. mol. %	Na_2O Wt. %	Sericitization
ÖA 16	0.4	1.47	5.0	3.02	Generally strong
17	0.6	1.36	4.5	3.01	Patchy, moderate
18	0.2	1.86	4.5	2.93	Strong
19	0.2	1.68	5.4	2.94	Strong
20	—	1.63	5.1	2.59	Strong
21		1.58	4.6	3.64	Patchy, locally v.strong

Al may be slightly deficient nearest the vein as compared with the more distant samples, but the data are not absolutely conclusive. No consistant change of Si or Mg + Fe can be demonstrated.

The calculated cation percents of the samples from the amphibolite band yielded curves with little variation. It was considered possible that the rise of K within 2 m of the vein was real, being due to the increase of biotite near the vein. However, the irregular variation in the degree of sericitization, which was impossible to estimate quantitatively, made any conclusion hazardous. K_2O and Na_2O were therefore determined with the flame photometer. The weight percents determined and the cation percents which were calculated are compared in table 6. The agreement is not good and the changes determined are irregular, but all changes are very small — too small to be accorded any significance.

The change in the mineralogy and the chemistry of the surrounding rocks in the vicinity of the vein may have taken place with or without a change of volume of the rock. In order to determine this the volume change calculation which was introduced in a previous publication (REITAN, 13, p. 306) was used. In this calculation the relative volume occupied by a standard number of oxygen ions is determined after having determined the mol volumes of the minerals of the rocks (one mol of each mineral being defined on the basis of the formula containing 24 oxygen ions).

In the five gneiss series' and in the amphibolite series the sample nearest the vein was compared with the average of the rest of the series. The percent change in going from the average to the rock

Table 7

Calculated volume change

Sample no.	Volume no.	Difference	% change nearest vein
ÖG 27 .... Av. ÖG 28–33	29796.5 29178.3	618.2	+ 2.12
ÖG 34 .... Av. ÖG 35–38	29301.9 28894.0	407.9	+ 1.41
ÖG 39 .... Av. ÖG 40–43	29674.0 29331.3	342.7	+ 1.17
ÖG 44 .... Av. ÖG 45–47	29781.6 29073.5	708.1	+ 2.44
ÖG 48 .... Av. ÖG 49–51	29934.1 29042.3	891.8	+ 3.06
ÖA 16............ Av. ÖA 17–21........	28851.3 28754.6	96.7	+ 0.34

nearest the vein is shown in table 7. In all cases there was an increase of volume nearest the vein. The increase is much larger for the gneiss than for the amphibolite. The significance of this conclusion is more appropriately discussed in connection with the genesis of the pegmatite; further comment is deferred to p. 219.

Genesis of the pegmatite vein

So far virtually no mention has been made of the structure in the area. Little can be said. The pegmatite vein cross-cuts the parallel structure of the gneiss and the amphibolite bands. In almost all cases an amphibolite band which is cut off on one side can be found again on the other side of the vein. There has been no apparent off-set of the bands; dilation, if it occurred, must have been parallel to the strike. However, only one set of parallel structures crosses the pegmatite vein and therefore no conclusive answer to whether or not there has been dilation can be given on structural grounds.

The pegmatite cross-cuts the amphibolite bands and the parallel structure of the gneiss and amputates an inter-boudin pegmatite and a white, plagioclase pegmatite vein filling in an amphibolite (cf. map, fig. 1). It must therefore be considered to be later than these rocks. As the white, plagioclase pegmatites (both the inter-boudin pegmatites and the vein pegmatites) are a product of the main dynamic metamorphism of the area, the pink pegmatite is later than the main deformation of the area.

To decide upon the mechanism by which the pegmatite formed, the following questions must be answered. Did the material which now constitutes the pegmatite come from the immediately surrounding rock or from elsewhere? Is the pegmatite a replacement pegmatite or did it form in a crack (dilation of the crack being parallel to the strike)?

The reason for taking the series' of samples and for doing the quantitative petrography was to be able to answer the first question. The curves obtained for the quantitative variations of the mineralogy of the rock surrounding the pegmatite, and even more clearly and strikingly the curves obtained when the calculated cation percents are plotted against distance from the vein, show that there is a relative increase of K and a relative decrease of Na, Ca, and possibly Al near the vein. The pegmatite vein consists predominantly of microcline with some quartz and plagioclase and a lesser amount of biotite and hornblende. This means that the cation which (relative to the average gneiss) is most enriched in the pegmatite vein is also most enriched in the immediately adjacent gneiss and diminishes with distance from the vein; those elements which (again relative to the average gneiss) are most deficient in the pegmatite (namely Na, Ca, and to a lesser extent Al) are deficient in the immediately adjacent gneiss and increase with distance from the vein.

This corresponds to the situation described graphically in fig. 14a and c. With respect to one cation (in this case K), enrichment in the vein and relative enrichment of that cation in the adjacent rock which diminishes with distance from the vein, implies diffusion of that element from its site of greatest enrichment (the vein) into the surroundings. With respect to another cation (Na, Ca) which is relatively deficient in the pegmatite and in the immediately surrounding rock but less deficient at some distance from the vein, movement of another cation into the surrounding rocks is implied such that the cation concerned is relatively depleted in the surrounding rocks. This

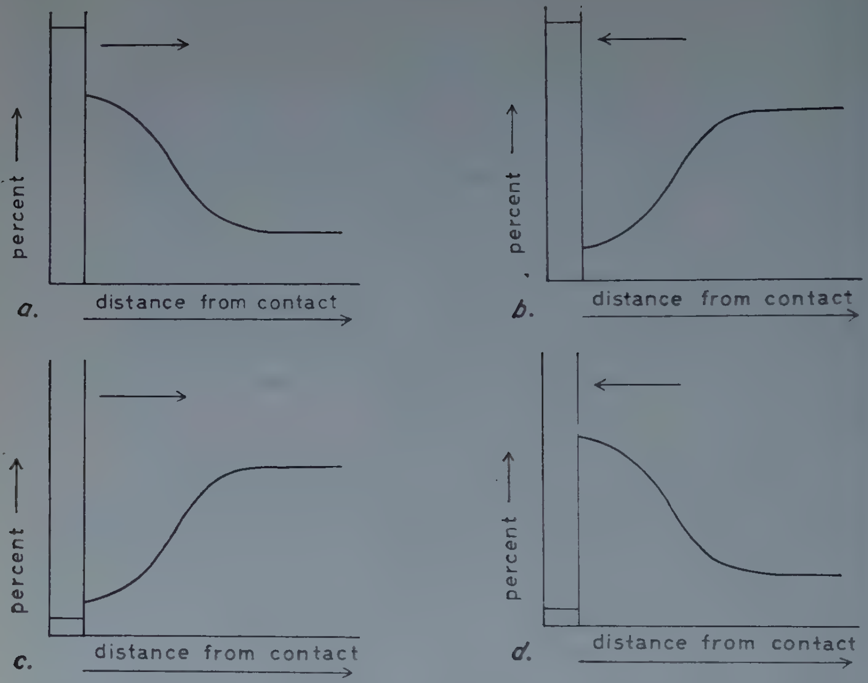

Fig. 14. Theoretical curves for the variation of a component in a pegmatite vein and with distance from the vein in the surrounding rock. a. Concentration high in the vein and diminishing with distance from the vein in the surrounding rock implies movement of the component from the vein into the surroundings. b. Concentration high in the vein and increasing with distance from the vein in the surrounding rock implies movement of the component from the surrounding rock into the vein, i.e., enrichment in the vein at the expence of the surroundings. c. Concentration low in the vein and increasing with distance from the vein in the surrounding rock — the complement of a. — implies movement of another (other) component(s) from the vein into the surroundings. d. Concentration low in the vein and decreasing with distance from the vein in the surrounding rock — the complement of b. — implies movement of another (other) component(s) from the surrounding rock into the vein; i.e., residual enrichment of the component near the vein.

is the complement of the former. Fig. 14b and d illustrate the case in which cations move from the surrounding rock into the vein. In the first case, a cation is «leached» from the surrounding rock and enriched in the vein. In the second case, a cation which does not move into the vein and is deficient there is residually enriched in the surrounding rock as another (other) cation(s) moves into the vein.

It is readily seen that the curves obtained for the gneiss at the western end of St. Hansholmen correspond to those shown in fig. 14, a and c. Therefore, material (primarily K) has moved from the vein into the surrounding rock.

This answers the first question. The material which now constitutes the vein did not come from the immediately surrounding rock. The material came from elsewhere and excess material moved out from the vein into the surrounding rocks.

One asks, from where did the material come?

The most obvious source and the most probable one is the group of granulite facies rocks, migmatites, banded gneiss etc., which according to the map of the facies series of the Arendal district (Bugge, 3, fig. 26, p. 134) occur immediately to the southwest of St. Hansholmen. If the material of the rocks constituting St. Hansholmen, the mineral assemblage of which corresponds to the amphibolite facies, was, at the time of the formation of the pegmatite vein, not sufficiently mobile to move into the zone in which the pegmatite formed,[1] then the most likely place from which the material could have been derived is from the granulite facies area nearby, either below (?) or laterally.

It is therefore believed that a K-rich phase derived from the granulite facies rocks in the vicinity, constituted the material from which the pegmatite vein formed, and that some of this material also moved into the rocks surrounding the vein and led to the presently observable relative K enrichment and Na and Ca depletion.

There is no direct evidence as to whether the vein is a replacement vein or a dilation vein. However, the volume change calculations (p. 216) have shown that in every series of samples there has been a relative increase in volume of the rock close to the vein as compared with that more distant. This is of course to be expected if a crack had formed. The following sequence of events can be envisaged. A crack occurred in the banded gneiss. As a result the pressure in the surrounding rock was lowered in the direction perpendicular to the vein and while the crack was being filled by newly crystallized minerals from the phase introduced into the crack a small amount of expansion took place in the surrounding rock.[1] As the adjacent rock expanded after the release of pressure, material from the vein moved out into the surrounding rocks and there partly replaced the original minerals but also crystallized between the grains where the pressure was now low. Expansion in the surrounding rock would cease after the vein had been filled with crystalline material; diffusion of material from the vein into the surrounding rocks would stop as no new low pressure sites would be created. Note that the expansion in the amphibolite band was very little, while that in the gneiss was very much more. This is what would be expected on the basis of what is known about the relative rigidity of the rocks. The lack of chemical change in the amphibolite adjacent to the vein also indicates that it could not «blot up» material from the vein to the same extent as could the gneiss; there were very few low pressure sites in the amphibolite which were available to filling by material from the vein.

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[1] Note that as yet no standpoint has been taken as to the kind of medium from which the pegmatite formed or what form — magma, solution, suspension, or diffuse state — the material had which moved to the site now occupied by the pegmatite. This is not necessary at this point in the discussion.

The lack of structural deformation associated with the cross-cutting vein would suggest that if it is a replacement vein it is most likely a volume-for-volume replacement vein. If this is the case, it is difficult to understand why the adjacent rocks have undergone a volume increase near the vein. The lack of deformation in the banded gneiss near the vein (which can be associated in time with this vein) hardly allows for the proposal that the vein is a replacement vein in a fold — or shear — determined low pressure zone crossing the banded gneiss.

It seems, therefore, most probable that the vein formed in a crack in the banded gneiss and that the material which filled the vein was derived from the rocks of the nearby granulite facies zone.

The final question to be answered in connection with the genesis of the pegmatite is: What was the nature of the pegmatite forming fluid?

Four possibilities must be considered. They are: 1) a concentrated fluid consisting primarily of the silicate components of which the pegmatite is now composed, in which a certain amount of water may have been dissolved, i.e. a magma; 2) a watery fluid in which the silicate materials now constituting the pegmatite were dissolved and which carried the dissolved components to the present site of the pegmatite where they precipitated, i.e. a hydrothermal solution; 3) a fluidized system, i.e. a gas-solid system consisting of a suspension of solid particles in a through-flowing gas (REYNOLDS, 11); and 4) a disperse phase consisting of the components of the pegmatite in ionic, molecular, or atomic form, accompanied by water also in disperse form, which diffused to the site of the pegmatite and there crystallized.

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[1] The gneiss was apparently brittle enough to crack and sufficiently inert at the time of the formation of the pegmatite that it contributed no or extremely little material to the phase entering the crack — additional evidence that the vein formed quite late in the metamorphic history of the area.

Possibility number two is discarded as quite unlikely for the following reasons. Large quantities of water would be necessary in order to dissolve and transport the material which now appears as the pegmatite vein. Calculations based on the solubilities which Friedman (4) arrived at in his study of the system $H_2O-Na_2O-SiO_2-Al_2O_3$ have led to the estimate that for every cm^3 of pegmatite material deposited, ca. 2×10^2 cm^3 of solution would have had to have passed the site of the pegmatite (REITAN, 12). SVERDRUP (15), basing his calculations on the solubilities of quartz and albite in superheated steam given by MOREY and HESSELGESSER (8), and assuming that *all* of the dissolved material precipitates at the site of the pegmatite, has arrived at approximately the same figure (1.91×10^2 cm^3 of solution yielding 1 cm^3 of pegmatite material).

However, the possibility that the pegmatite has been precipitated from a hydrothermal solution becomes even more unlikely when one attempts to calculate the amount of rock which would have to be «dehydrated» in order to provide the hydrothermal solution.

It has been suggested as most likely (p. 219) that the material which now constitutes the pegmatite vein was derived from the granulite facies rocks nearby. As no granite bodies are close to St. Hansholmen, the most reasonable source from which a hydrothermal solution could be derived is these granulite facies rocks. Such rocks are generally quite «dry», but it is just that water which has been expelled from them which could conceivably have constituted a hydrothermal solution which bore the material now constituting the vein to its present site.

If we assume that from 10^2 cm^3 of the rock one derives 10 cm^3 of solution[1] then, assuming the solubility previously arrived at (ca. $2 \times$

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[1] None of the rocks in this area are particularly «wet», and as the formation of the vein occurred late in the main metamorphism of the area it seems unlikely that the rocks of the granulite facies would lose more than 1 % H_2O.

10^2 cm^3 of solution being capable of yielding 1 cm^3 of solid pegmatite material), ca. 2×10^4 cm^3 of rock yield enough solution to form 1 cm^3 of pegmatite. If we assume that the vein on St. Hansholmen, which has a surface area of about 20 m^2, was in all 500 m in vertical extent, then 0.4 km^3 of rock must have been «dehydrated» to yield solution sufficient to deposit this quantity of pegmatite. However, the vein exposed on St. Hansholmen is probably much longer than the exposure on the island. On the adjacent islands of Lille Stangholmen, Kaninholmen, and Engholmen, identical veins can be seen and these exposures all lie essentially in line with one another; they are undoubtedly parts of the same vein (personal observations, see map, fig. 3, p. 227 in REITAN, 12, for the location of these islands). If these are all parts of the same vein then the length of the vein is at least 500 m. If the width is assumed to be constant at 2 m and the vertical extent taken to be 500 m, then 10 km^3 of rock would have had to have been «dehydrated» in order to yield the quantity of hydrothermal solution necessary to form the vein.

That the pegmatite vein has crystallized from a fluidized gas-solid system is a possibility, but none of the characteristic features of veins for which REYNOLDS (11) has suggested such an explanation are to be seen in the pegmatite vein on St. Hansholmen. The explanation has its appeal in that it could explain the vein as the result of minimal or no dilation while yet allowing for some expansion of the surrounding rocks. However, as positive evidence in its favor is completely lacking it will not be advocated in this case.

To decide between the two remaining possibilities is difficult. Temperature determinations using Barth's feldspar thermometer (2) have been made, taking co-existing feldspars both from samples of the gneiss and of the pegmatite. The results are shown in table 8. The temperature of the pegmatite is considerably higher than that of the gneiss. The differences between the temperatures determined at three points in the pegmatite are so small that no importance can be attached to them. (For example, if the plagioclase in ÖP17 is Ab_{74} instead of Ab_{73} as determined, and that in ÖP18 is Ab_{68} instead of Ab_{69}, the temperature determined would be the same.) However, the difference between the pegmatite vein and the gneiss is significant, being about 100° C.

The temperatures determined in the gneiss are irregularly variable,

Table 8

Temperatures indicated by co-existing feldspars

Sample no.	ÖG27	ÖG30	ÖG31	ÖG32	ÖG33	ÖG39	ÖG40	ÖG41	ÖG42	ÖG43	ÖG44	ÖG45	ÖG47	ÖG48	ÖP17	ÖP18	ÖP19
Microcline																	
Na_2O	1.24	1.30	1.64	1.60	1.32	0.98	0.84	1.10	0.94	1.01	1.07	1.25	1.08	1.24	2.40	2.50	2.09
K_2O	14.36	13.44	14.24	13.88	14.75	15.45	15.42	15.08	15.25	15.00	15.00	14.57	14.63	14.85	12.96	13.50	13.27
Mol. prop.																	
ab.	11.6	12.8	14.9	14.9	12.0	8.8	7.7	10.8	8.6	9.3	9.8	11.5	10.2	11.3	22.0	20.2	19.3
or	88.4	87.2	85.1	85.1	88.0	91.2	92.3	90.0	91.4	90.7	90.2	88.5	89.8	88.7	78.0	79.8	80.7
Ab-content of associated plagioclase .	80	70	76	71	70	75	72	71	70	71	80	72	71	80	73	69	69
k.	0.14	0.18	0.20	0.21	0.17	0.12	0.11	0.14	0.12	0.13	0.12	0.16	0.14	0.14	0.30	0.29	0.28
Temperature	410	460	480	490	450	390	380	410	390	400	390	430	410	410	570	560	555

tending to be lower near the vein. This is probably because equilibrium was not established. The textures displayed by the microcline in the gneiss indicate that this is the case; had the plagioclase and microcline been in equilibrium one would not expect that the microcline would be present as intergranular stringers. If equilibrium was not established the tendency towards lower determined temperatures near the vein is easily explained.

A K-rich phase was introduced into the vein and this phase became more Na-rich as crystallization in the vein proceeded (cf. the change of composition of the feldspars in the vein, p. 200). Consequently the early metasomatism of the surrounding gneiss was very predominantly a K-metasomatism. Naturally the first material introduced into the gneiss crystallized at the first suitable low pressure sites arrived at, i.e. closest to the vein. Thus the mol fraction of albite in the alkali feldspar (the term which appears in the numerator of the distribution equation) is much lower than would be the case were equilibrium achieved. Material which enters into the gneiss later would be more Na-rich, but would be forced to penetrate further into the gneiss as suitable low pressure sites near the vein would have been already occupied (at least the percentage of late material would be higher further from the vein). Still another factor tending in the same direction is that the proportion of second generation microcline is highest nearest the vein. First generation microcline is probably in equilibrium with the coexisting plagioclase, and where this predominates the temperatures are higher and probably more nearly correct; the temperature determined for the gneiss at the opposite end of the island is 460° C. However, as equilibrium was not attained near the vein the temperatures determined in the gneiss near the vein have no significance and inconsistent and irregular variations are only to be expected.

The thin, more acidic (?) rims frequent on plagioclase adjacent to microcline also indicate that equilibrium was not achieved. They might be taken to indicate that the feldspars began to react such that a more acidic plagioclase began to be formed; the reaction which would take place with lowering of temperature. That this reaction was only initiated near the vein could be because only there were there sufficient volatiles to facilitate the reaction. If these rims are the result of limited reaction at lower temperature they indicate that the surrounding rock was not appreciably heated near the vein at the time of its formation, and support the contention that the vein is late in the metamorphic history of the area.

Neither are the triclinicity measurements of any help. The triclinicity of the microcline in the pegmatite is higher than that in the gneiss (see table 2, p. 208), but this may be accounted for by the cooling history of the microcline at the two places and the probable abundance of interstitial fluids which would assist in the diffusion necessary to accomplish the reconstructive transformation which occurs in the ordering of microcline. The microcline in the gneiss was largely derived from the same phase as that in the pegmatite. However, the microcline in the gneiss crystallized at points where the surroundings were cooler than the vein. As equilibrium between the microcline and plagioclase was not achieved in the gneiss, one must assume that the minerals in the gneiss were rather passive at the time and since the microcline formed, this passivity probably being due to relatively low temperature. Thus, microcline formed in the gneiss, probably originally quite poorly ordered (GOLDSMITH, 5; GOLDSMITH and LAVES, 6), would have had less opportunity to achieve order than would the microcline in the pegmatite which cooled more slowly.

The possibility that the H_2O pressure was higher in the vein is also good in that the material which constitutes the vein must have been accompanied by considerable H_2O, no matter what the nature of the pegmatite forming fluid. If the fluid was a magma it must have been under rather high H_2O pressure in order to be mobile at a temperature below 600° C, and, as H_2O is among the most fugitive components of rocks, a disperse phase must also have contained considerable H_2O. The presence of H_2O (and other gases?) would assist in the diffusion necessary to order the Si and Al in the microcline.

The changes in the composition of the feldspars from the vein boundaries and towards the center are those which would be expected had the pegmatite crystallized from a magma. Both feldspars become more rich in Na towards the center of the vein, which is assumed to be latest.

If the vein crystallized from a disperse phase which migrated from the granulite facies rocks through the crack in which the pegmatite formed, this change of composition of the plagioclase may or may not follow naturally. It would appear that Na diffuses more rapidly towards the zones of low pressure than Ca (RAMBERG, 10; REITAN, 12, 13, and 14). One should, therefore, predict that the plagioclase formed first would be richer in Na than that which formed later, as was the case in the veins in the olivine gabbro near Risør (REITAN, 13). This is not the case. However, at high temperatures, K metasomatism is much more predominant than Na metasomatism (RAMBERG, 9, p. 206), and might imply higher mobility of K than of Na at the temperature of formation of the vein on St. Hansholmen. If such is the case one would predict a more K-rich alkali feldspar first, followed by more Na-rich alkali feldspar. This is the case. But if the distribution of Na between the two feldspars is controlled by the temperature of their formation and the temperature is essentially constant from the first to the last, the feldspars produced will be a function of the limiting quantities of the various components. Thus, if the quantity of Na in the environment is the limiting factor, the composition of the feldspars will be controlled by its distribution. If the primary change in the environment from which the feldspars crystallized was an increase in Na, such that the alkali feldspars crystallizing out through time were forced to include more Na, then, at constant temperature, the plagioclase formed must also contain more Na if equilibrium is to be maintained (BARTH, 2). In other words, the composition of feldspars produced will be a function of the temperature and bulk composition of the phase from which they form. If formed from a magma the liquid must become more Na-rich through time and both feldspars become more Na-rich. If formed from a diffuse phase they may become richer or poorer in Na, depending on whether the controlling change in the composition of the environment is of Na or Ca or K.

The temperature at which the pegmatite formed — ca. 560°-570° C — is apparently not too low for the existence of certain magmas (i.e., molten silicate material). JAHNS and BURNHAM (7) report that the beginning of melting of the Harding, New Mexico, pegmatite under 2000 bars H_2O pressure occurred at 555 ± 5° C. The experimental work of TUTTLE and BOWEN (16) on the Quincy granite, Westerly granite, and the «ternary» (Ab-Or-SiO_2) minimum has shown that the beginning of melting curve at various water-vapor pressures approaches 650° C asymptotically as the pressure increases, at least up to 4000 kg/cm^2 pressure. However, further studies, with the proportion of alkalis in excess of the alkali-alumina ratio in alkali feldspar, have indicated that liquids rich in alkali silicates (and also water) can exist at lower temperatures. It is therefore impossible to dismiss a magma from consideration on the basis of the temperature determined for the crystallization of the pegmatite.

Tuttle and Bowen have also arrived at the conclusion that «at least some of the amphiboles may well be unstable in the presence of a moderate or high water-vapor pressure». (TUTTLE and BOWEN, 16, p. 92).

If the presence of hornblende in the pegmatite on St. Hansholmen requires the assumption that the water-vapor pressure was low, then the possibility that the pegmatite vein crystallized from a magma is reduced. At least moderate (ca. 1000 atm.) and probably higher water-vapor pressure must certainly be necessary (even if other volatiles are present) to allow the existence of a magma derived from ordinary rocks at temperatures of 650° C or below. (It is assumed here that a possible magma derived from the granulite facies rocks late in the metamorphic history of the region would have been a palingenetic magma resulting from differential melting of these rocks and «sweating out» of the molten material).

In conclusion it must be said that a definite answer to the final question — what was the nature of the pegmatite forming fluid? — cannot be given. There have been found neither features which decisively prove what the nature of the fluid was nor features which are fatal to all but one conceivable possibility. The question must remain unanswered.

Only two possibilities seem to exist. Either the material now constituting the vein crystallized from a magma which was injected into a crack formed in the banded gneiss, or it crystallized from a disperse phase composed of high energy particles (ions, molecules, atoms) which diffused (and in a sense also were injected) into a crack formed in the banded gneiss. Most readers will probably prefer to believe in one or the other, but the author is aware of no evidence which justifies a conclusion.

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Dr. H. Bjørlykke, director of the Geological Survey of Norway, has kindly allowed the publication of this paper and has allowed me to engage in this study while in the employ of the Geological Survey.

He has also made available to me the facilities of the Geological Survey of Norway. Thanks are also due to Miss K. Haugen who has performed almost all of the alkali determinations. The staff of the Geological Survey and the Mineralogisk-Geologisk Museum, University of Oslo, have at all times willingly given their assistance when it was needed. Dr. K. Heier has often discussed the problems of this study and related subjects with me, and for his time and interest I extend my thanks. For making available the facilities of the Mineralogisk-Geologisk Museum, and for very valuable criticism and advice and his unfailing interest and encouragement I wish especially to thank Prof. T. F. W. Barth.

Errata

In Pegmatite Veins and the Surrounding Rocks: I. Petrography and Structure : Norsk Geologisk Tidsskrift, Bind 36, Hefte 3, pp. 213—239, 1956, the composition of the plagioclase in sample ÖP17 was incorrectly reported. The following changes should be made.

p. 221 (in two places) — for An_{35} read An_{27}.

This change results in a change in the determined temperature of the pegmatite vein.

p. 235 — for 610° C read 570° C.

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CONTRIBUTIONS TO THE MINERALOGY OF NORWAY

No. 1. An Introduction.

BY

HENRICH NEUMANN

Apart from more voluminous monographs the literature on the mineralogy of a certain area tends to consist of a great number of short papers and even of mere notes putting on record finds of a rare mineral or of a certain mineral in an interesting or unusual environment. As far as Norway is concerned such small papers and notes have also in part been published in journals which are not too widely circulated and read, and they are therefore easily overlooked even by Scandinavian mineralogists, not to mention the international group of interested readers.

This new series "Contributions to the Mineralogy of Norway" is introduced as an attempt to make the future literature on Norwegian mineralogy more conveniently available and more easily reviewed. The "Contributions" are not intended to be the "house bulletin" of any one institution nor a convenient tool for any one group of coo-perating scientists, on the contrary any mineralogist working within the field concerned, Norwegian or otherwise, is urged to submit his papers for publication in this series, thereby helping it to serve its purpose.

In 1948 I. Oftedal published a review of minerals found in Norway up to that time (Oversikt over Norges mineraler. Norges Geologiske Undersøkelse No. 170 (1948)) with short comments on their mode of occurrence. It would seem appropriate in an introduction to the "Contributions" to add to Oftedal's list of minerals a few species which were overlooked at the time, and also to list those species which have been found in this country since 1948:

Allophane from a vein through lower Ordovician black shale in Oslo. (30).[1]

Ancylite from Langesundsfjord. (G.M.)[2]

Anglesite from Kvaløy, as an alteration product of galena. (Found by T. Sverdrup).

Annabergite from Råna nickel deposit as an alteration product of niccolite. (G.M.)

Bastnäsite see the following paper of this journal.

Bavenite from a pegmatite dyke at Boksjøen near Aspedammen in the county of Østfold as a younger reaction product of beryl. (G.M.) (Found by O. J. Adamson).

Birnessite from supergene infillings of manganese minerals in fissures in nordmarkite at Lillevann near Oslo. Birnessite may not be a valid species. (G.M.)

Brannerite in open fissures in green schist near Åmot in Modum and probably genetically related to granitic pegmatite dykes near by. (4)

Brookite from a pegmatite dyke on Straumsheia. (24).

Cahnite from Klodeborg mine at Arendal (7).

Carrolite from Borras copper deposit at Alta in the county of Finnmark. (53).

Chamosite from Kongsberg, Norway.[3]

Clausthalite from Kongsberg and Nummedal. (13).

Cookeite from a pegmatite dyke at Holandsfjord, N. Norway. (G.M.)

Davidite from the Tuftan pegmatite dyke, Iveland, Setesdal. (G.M.)

Epsomite from Storhelleren, Fauske in the county of Nordland. (G.M.)

Euclase from Iveland as an alteration product of beryl. (50).

Gahnite from Tor's mine, Øvre Vats in a pegmatite dyke.[4] (Found by H. Ramberg).

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[1] Numbers in brackets refer to literature list at the end of this paper.

[2] Species found and/or identified by members of the staff of the Geological Museum of Oslo University are marked G.M. Further details will be published later.

[3] See G. W. Brindley: X-ray Identification and Crystal Structures of Clay Minerals. London 1951. Table II, 7 facing p. 72.

[4] See R. C. Sørbye: Geological Studies in the North-Eastern Part of the Haugesund Peninsula, Western Norway. Univ. i Bergen, Årbok 1948, Naturv. Rekke No. 6. (See p. 59).

Gonnardite (= ranite) from Langesundsfjord. (19).

Halite from a drill core of nepheline syenite pegmatite at Bekkarfjordnes on the island of Seiland in the county of Finnmark. (G.M.)

Hisingerite from pegmatite dykes at Hyttåsen, Mjærum farm in the county of Østfold and at Drag, Tysfjord in the county of Nordland. (G.M.)

Jarosite as a secondary mineral on graphite schists from several localities. (G.M.)

Kasolite from a pegmatite dyke on Straumsheia as an alteration product of uraninite. (24).

Lanthanite from small cavities in the Seteråsen pegmatite dyke on Mjærum farm, Hobøl in the county of Østfold. (G.M.)

Lizardite from a serpentine-magnesite deposit in Snarum.[1]

Margarite from eclogites and labradorites in Sunnmøre.[2]

Mesolite from skarn iron ore deposits at Arendal. (8).

Neodigenite from the Ulveryggen deposits, west of Repparfjordbotn in the county of Finnmak. (52).

Palygorskite from Åna Sira (3) and from Rømteland. (G.M.)

Paragonite from an acid dyke cutting through an ultrabasic rock at Bruvann, Råna in the county of Nordland, and in mica schists with cyanite at Rognan, Saltdalen in the county of Nordland. (G.M.)

Pyroxmangite (= sobralite) from sedimentary iron-manganese ores in Ofoten, Northern Norway.[3]

Ramsdellite from Lian manganese deposit near Mandal. (G.M.)

Scolezite from skarn iron ore deposits at Arendal. (8).

Siegenite from Raipas mine at Alta in the county of Finnmark. (53).

Stoltzite from a pegmatite dyke on Foss farm in Sigdal. (G.M.)

Synchysite from Kongsberg. (G.M.)

Tysonite from pegmatite dykes at Jennyhaugen, Drag and Hundholmen, both in Tysfjord in the county of Nordland. (Found by T. Sverdrup).

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[3] S. Foslie: Håfjeldsmulden i Ofoten og dens sedimentære jernmanganmalmer. Norges Geologiske Undersøkelse No. 174 (1949). See p. 56.

Uvarovite from a chromite deposit near Røros, and from Kalkovnen, Grua near Oslo.

Yttrialite from the Ivedal pegmatite dyke in Iveland, Setesdal. (G.M.)

The following species described from Norway have been discarded since 1948:

Apoanalcite (31)
Ranite (= gonnardite) (19)
Weibyeite (G.M.)

In the paper mentioned above I. Oftedal gives a list of literature references with 136 entries on Norwegian minerals and mineral deposits. The list is not by any means complete but together with literature referred to in the papers listed by Oftedal it will cover the field fairly well. Below follows a list of Norwegian mineralogical publications printed since 1948.

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CONTRIBUTIONS TO THE MINERALOGY OF NORWAY

No. 2. Bastnäsite, a new mineral for Norway.

By

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Abstract: Optical and X-ray powder data are given for bastnäsite from four localities in Norway. The indices of refraction and the birefringences are lower than those reported earlier. The crystallographic constants are in close agreement with the previously published values, except in one case where the mineral shows smaller a_0 and c_0 values. The variation in the unit cell and the optical properties are discussed, but lack of chemical data prevents any conclusive explanation. Bastnäsite has been found as a secondary mineral, mostly as an alteration product of allanite. The mineral seems to be more widespread than supposed earlier, especially in pegmatites containing altered allanite.

Introduction.

Bastnäsite, which in recent years has been reported from several localities in the world, has not previously been found in Norway. During the last 1½ years the mineral has been found at four localities, three in southern Norway and one in northern Norway.[4] The localities, which are plotted in fig. 1, are:

1. The Rømteland pegmatite, Vest-Agder county, Southern Norway. (SVERDRUP, 1957).
2. The nepheline-syenite pegmatite, Øvre Arø, Langesundsfjord, Telemark county, Southern Norway. (BRØGGER, 1890, II).
3. The Mostad pegmatite, Aust-Agder county, Southern Norway.

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[1] Norges geologiske undersøkelse, Oslo.

[2] Norges geologiske undersøkelse, Oslo.

[3] Mineralogisk-Geologisk Museum, University of Oslo.

[4] After ending this work Sverdrup and Sæbø have identified bastnäsite from three pegmatites in the Tysfjord district, Northern Norway. These finds will be reported later.

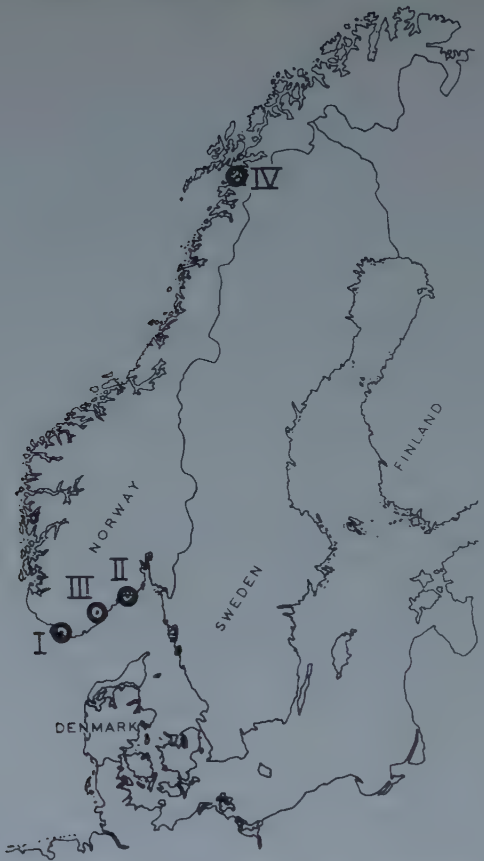

Fig. 1. Map showing the situation of the bastnäsite localities.

4. The Nedre Lapplægret pegmatite, Drag in Tysfjord, Nordland county, Northern Norway.

The main part of this work has been carried out at the X-ray laboratory, Mineralogisk-Geologisk Museum, University of Oslo. All X-ray work has been done with 9 cm cameras and filtered iron radiation, 14 mA, 40 kV.

ACKNOWLEDGEMENTS.

We are indebted to Prof. T. F. W. Barth and Dr. H. Neumann for having discussed some of these problems with us, and for their interest in and encouragement for our work. Mr. P. H. Reitan has corrected the text and checked some of our observations on the optical properties, and Miss B. Mauritz has taken the photographs.

Bastnäsite.

Bastnäsite has the general formula R ((F) (CO_3)) where R=Ce, La, Dy etc. (KLOCKMANN-RAMDOHR, 1954). The structure was investigated by OFTEDAL (1929, 1931) and DONNAY & DONNAY (1953). They confirmed the hexagonal symmetry and found that the space group is ($C\bar{6}2c$). Donnay & Donnay also showed the relations to the minerals parisite, røntgenite and synchisite.

1. *Bastnäsite from the Römteland pegmatite, Vest-Agder county.*

Geology. This is the first locality for bastnäsite in Norway. For a detailed description of this pegmatite see SVERDRUP (1957).

From the host rock (farsundite (BARTH, 1928)) to the centre of the pegmatite the following zones are observed; a graphic granite zone, a mineralized zone, a quartz-feldspar zone, and a central quartz mass. Within the pegmatite some lumps of amphibolite occur. They are always bordered with a strongly mineralized cerium-rich zone, different from the ordinary mineralized zone in the pegmatite containing Y-bearing minerals (SVERDRUP 1957). Bastnäsite is found in the mineralized zone around the amphibolite lumps as an alteration product of allanite. (Bastnäsite as an alteration product of allanite was mentioned from Bastnäs, Sweden, by NORDENSKJÖLD, 1868) It occurs as a very fine grained impregnation in the altered, brown to dark brown metamict rim zones of the allanite crystals.

Optics. It has not been possible to make any reliable determinations of the refractive indices because the only grains available are too small. It seems, however, to be optically (+). (See table no. 1).

X-ray data. The mineral has been identified by the X-ray powder pattern (see fig. 5). The crystallographic constants are:

$a_o = 4.08$ and $c_o = 4.89$. Axial ratio $c/a = 1.383_5$. (See table no. 1).

2. *Bastnäsite from Øvre Arø, Langesundsfjord, Telemark county.*

Geology. The geology and mineralogy of these famous mineral deposits have been dealt with by W. C. BRØGGER (1890, II).

Material. A sample, labelled "weibyeite", collected by W. C. Brøgger in 1902 from the eudidymite deposit (BRØGGER 1890, I and II) on the island of Øvre Arø, was found to contain bastnäsite. The

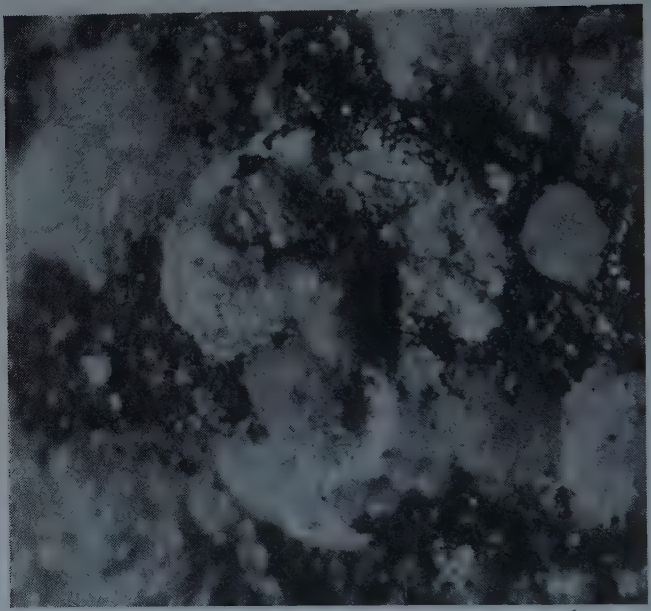

Fig. 2. Bastnäsite crystals from Øvre Arø, Langesundsfjord.
About 80 x.

mineral occurs as small grey-yellow to brown, short, hexagonal, barrel-shaped crystals in cracks and druses in the rock (Fig. 2). The crystals are up to 0.5 mm in thickness and 1 mm in diameter, but usually they are smaller. Some of the crystals show a marked difference between the central and the peripheral part. The centres of the bastnäsite crystals are yellowish and often quite dull; the rim zone, however, are glassy clear with the pearly-grey lustre typical of this mineral. In some places bastnäsite also occurs as a brown film on the surface of the other minerals.

Paragenesis. Bastnäsite is found together with crystals of analcite, aegirite, albite and small bipyramidal crystals of zircon. Weibyeite has not been found.[1]

Optics. Under the microscope the mineral shows a well-developed basal cleavage. The clear rim zone of the crystals often shows zoning parallel to the prism faces, the innermost zones often having higher indices of refraction than the peripheral parts. Sections parallel (001) show the uniaxial positive character of the mineral.

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[1] In his Mineralogische Tabellen, 3 Auflage, p. 425, Strunz says: «Weibyeite ? ähnlich Bastnäsite.», without giving further data. Strunz must therefore be regarded as the first to recognize bastnäsite in Norway.

The indices of refraction vary slightly, $1.691 < \omega < 1.700$; a few grains as low as 1.675; $1.780 < \varepsilon < 1.790$; $\varepsilon - \omega \approx 0.090$. These values are lower than usually found for bastnäsite and approach those found for parisite. This fact is, in our opinion, one of the reasons why BRØGGER (1890 II) came to the conclusion that this mineral probably was parisite; it is, of course, not excluded that parisite may occur in this paragenesis. Bastnäsite is only slightly attacked by strong HC1 and the X-ray powder pattern of the residuum still shows the pattern of bastnäsite. The optical constants remain unchanged.

X-ray data. The calculated crystallographic constants are: $a_o = 4.09$, $c_o = 4.90$, $c\ a = 1.383_4$ (table no. 1 and fig. 5.).

3. *Bastnäsite from the Mostad pegmatite, Aust-Agder county.*

Geology. The bastnäsite from this locality is found in a pegmatite situated about 4 km E of Gjerstad station, Aust-Agder county. The pegmatite occurs as a 7 m broad dike in the Precambrian gneiss; it strikes about N 170^g and dips 70^g E. From a stretch of boggy land which covers the pegmatite to the SE, the dike can be followed 8—10 m up a cliff until it is covered by forested ground.

Material. Two mineral fragments were sent to Norges geologiske undersøkelse by Chr. Mostad, Gjerstad. These pieces were given to the X-ray laboratory of the Museum for closer examination. The pieces, one consisting mostly of acid plagioclase and some monazite, and the other a piece of a monazite crystal, are both partly covered with black and red-brown coatings having a metallic lustre. The X-ray powder pattern showed the presence of pyrolusite (MnO_2). Also, yellow xenotime occurs as small square crystals in the monazite mass. Two small prisms of allanite penetrate the monazite. They are hardly visible with the unaided eye. The small red-brown specks occurring as infillings in the other minerals, preferentially replacing the plagioclase, often also occurring along the irregular border between the plagioclase and monazite, turned out to be partly bastnäsite and partly an unidentified metamict mineral.

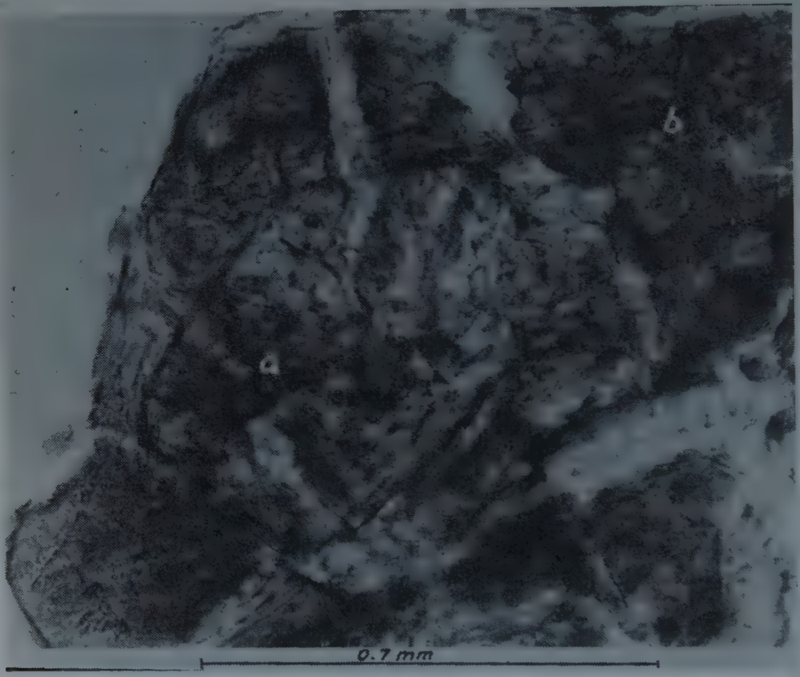

Fig. 3. Photo of the bastnäsite from the Mostad pegmatite. The polygonal bastnäsite crystal (a) inbedded in the metamict matrix (b) is marked by the distinct Becke-line. Section nearly parallel (001). Immersion liquid preparation in planepolarized light.

The bastnäsite is not well crystallized and on the X-ray powder pattern only the chief reflections can be seen. (See fig. 5).

Optics. Microscopical studies show that bastnäsite often as polygonal crystals, are embedded in a red-brown matrix (fig. 3.) that sometimes shows weak birefringence. The bastnäsite is uniaxial positive, and the index of refraction is considerably higher than that of the matrix, which was 1.670 ± 0.005. This indicates that the refractive indices of this bastnäsite lie in the range usually found for the mineral. Accurate optical data is impossible to obtain because the bastnäsite is always surrounded by the metamict mineral.

X-ray data. $c_o = 4.08$, $a_o = 4.89$, $c/a = 1.383_5$. (See table no. 1 and fig. 5).

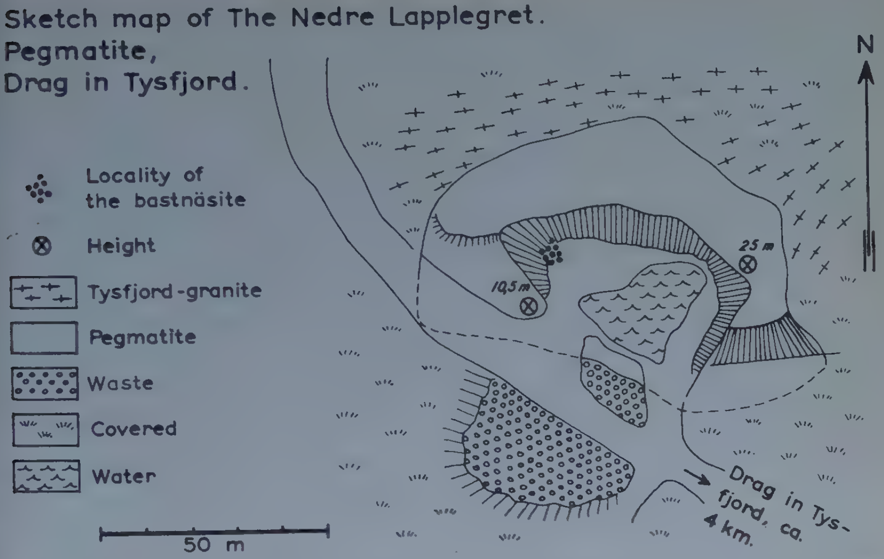

Fig. 4. Map showing the Nedre Lapplægret pegmatite. The dotted area shows the place at which the allanite crystals containing bastnäsite were found.

Genesis. The genesis of the bastnäsite seems to be rather complex. Its close association with monazite indicates that it is partly an alteration product of this mineral. Bastnäsite, however, is also found as traces (identified by means of the X-ray powder pattern) in one of the small prisms of partly metamict allanite.

The question of the genesis cannot be settled with the material available at present.

4. *Bastnäsite from the Nedre Lapplægret pegmatite, Drag in Tysfjord, Nordland county.*

Geology. This pegmatite is one of the biggest granite pegmatite bodies still in quartz-feldspar production at Drag in Tysfjord.

The pegmatite is formed like a thick, irregular lens partly conformable to the surrounding gneissic Tysfjord granite (see Foslie 1941, and Rekstad 1919). The pegmatite strikes about N 280^g. The pegmatite has been mapped and described by Sverdrup and Sæbø (1958). (See the map of the pegmatite, fig. 4.)

Material. Our material consists of some small pieces of the pegmatite (mostly quartz-feldspar aggregates) all containing several partly well formed crystals of allanite, 3—5 cm long and 0.5—1 cm broad. Sections perpendicular to (010) show rhombic outlines. Parts of the crystals are completely altered to yellow-red, powdery bastnäsite.

Optics. The material sometimes shows a finely laminated character, visible as pseudo-striation parallel to the long direction of the primary allanite crystal. The grains often also show a fine mesh structure that prevents even very small grains from showing any extinction at all. The bastnäsite is obviously mixed with metamict relics of the allanite. In one case a small single grain of bastnäsite showing a uniaxial positive interference figure, was found.

No reliable measurements of the indices of refraction were obtained; $1.691 < \omega < 1.700$ is probably correct. The value of ε is quite impossible to determine, and even a rough estimation cannot be given. (See table no. 1). This is disappointing in view of the crystallographic data obtained. The X-ray powder pattern gives a weak, but clear pattern of only bastnäsite. See fig. 5.

X-ray data. $a_o = 4.06$, $c_o = 4.86$ and $c/a = 1.350_4$. The cell size is definitely smaller than that of the other bastnäsites mentioned here, and also smaller than any reported in the literature (see table no. 1).

Paragenesis. Allanite and the secondary bastnäsite very often occur in direct contact with fluorite, which varies in colour from nearly colourless to the ordinary violet and pink. Otherwise, the specimens consist of white and pinkish microcline together with quartz and some albitic plagioclase. Muscovite and zircon are present as accessories.

Discussion.

In table no. 1 a_o, c_o, c/a and the optical data are given together with the data on bastnäsite and parisite of DONNAY & DONNAY (1953). Fig. 5 shows the reproduction of the X-ray powder patterns of the Norwegian bastnäsites, bastnäsite from Bastnäs, Sweden, and parisite from Halvorsrød, Råde, S. Norway.

The cell sizes of the bastnäsites from Øvre Arø, Langesundsfjord, the Römteland pegmatite, and the Mostad pegmatite are in good

Table 1.

Crystallographical and optical data.

Localities, ...	a_0	c_0	$a = a_0\sqrt{3}$	$c = 2c_0$	c/a	ε	ω	±
Bastnäsite, Rømteland	4.08	4.89	7.068	9.779	1.383_5	?	?	+
Bastnäsite, Øvre Arø, Langesundsfjorden ...	4.09	4.90	7.084	9.800	1.383_4	1.780– 1.790	(1.675) 1.691– 1.700	+
Bastnäsite, Mostad ...	4.08	4.89	7.068	9.779	1.383_5	much > 1.670		+
Bastnäsite, Lapplægret	4.06	4.86	7.033	9.497	1.350_4		1.691– 1.700	+
Bastnäsite[1]	4.13	4.89	7.16_2	9.78_7	1.36_7	1.815– 1.8242	1.713– 1.7225	+
Parisite[1]		4.67				1.765– 1.7712	1.670– 1.6742	+

[1] Donnay & Donnay, 1953.

agreement with each other and with earlier published data (Oftedal, 1931, and Donnay & Donnay, 1953). However, the mineral from N. Lapplægret has a definitely smaller cell size, and the relation c/a $= 1.350_4$ is also different from the others.

From table no. 1 it can be seen that on the basis of optical data it is impossible to say anything about the minerals from Rømteland and Lapplægret. The optical data for the mineral from the Mostad pegmatite are possibly in agreement with earlier described bastnäsites (Oftedal, 1931, and Donnay & Donnay, 1953).

The indices of refraction of the mineral from Øvre Arø are close to those of parisite (Donnay & Donnay, 1953). In general it is not possible to correlate the differences in cell sizes with the changes in the optical data for the Norwegian bastnäsites.

The X-ray powder patterns of the Norwegian bastnäsites are diffuse and weaker than the pattern of the original bastnäsite from Bastnäs, Sweden. This might indicate that Norwegian bastnäsites are in a state of incipient metamictisation. The low indices of refraction and low birefringence would seem to support this assumption.

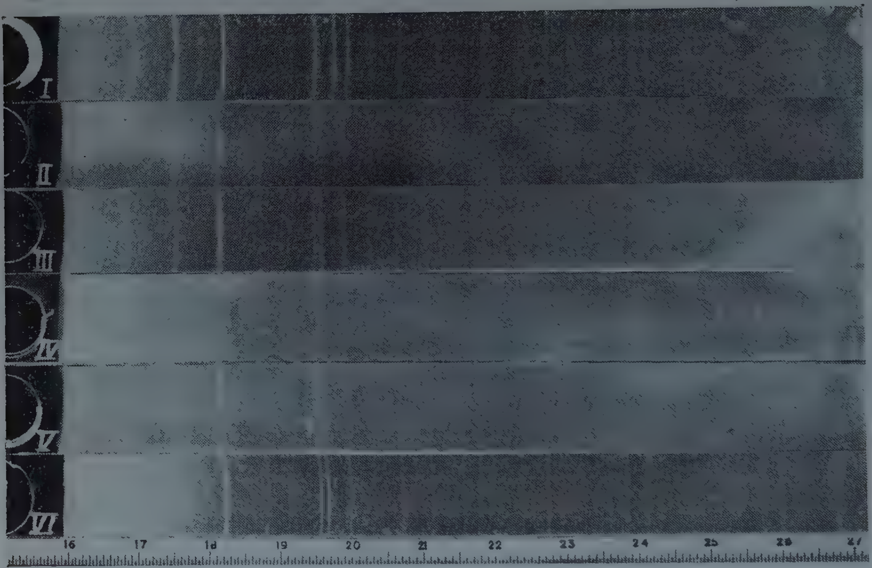

Fig. 5. Reproduction of the X-ray powder patterns.
I. Parisite, Halvorsrød, Råde, Southern Norway. Film no. 250.
II. Bastnäsite, the Rømteland pegmatite, Southern Norway. Film no. 9661.
III. Bastnäsite, Øvre Arø, Langesundsfjord, Southern Norway. Film no. 9725.
IV. Bastnäsite, the Mostad pegmatite, Southern Norway. Film no. 9605.
V. Bastnäsite, Nedre Lapplægret pegmatite, Northern Norway. Film no. 9705.
VI. Bastnäsite, Bastnäs, Sweden. Film no. 6247.

On the other hand, if OH-ions replace the F-ions in the structure the indices of refraction will decrease and the cell size will increase to some extent. This can be seen from the ionic radii: $OH^- = 1.40$ Å and $F^- = 1.36$ Å. This is probably the case in the bastnäsite from Ø. Arø. It is perhaps closer to the mineral kishtymite, which is thought to be a bastnäsite with OH^- instead of F^-. (Strunz, 1957.)

Conclusion.

The bastnäsite from the Mostad pegmatite is close to those described earlier (Donnay & Donnay, 1953). The same seems to be the case for the bastnäsite from Rømteland.

For the bastnäsite from the two other localities it is at present not possible to arrive at any conclusion as to the relations between the optical properties, crystallographic constants, and the chemical composition. We are, however, able to point out that the mineral bastnäsite shows a wide variation in its properties and we would like to stress the importance of using the X-ray powder pattern in its identification. Our finds of the mineral indicate that bastnäsite is a more common mineral than previously supposed, especially in pegmatites also containing altered allanite.

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FOSSILS(?) OF PRECAMBRIAN AGE FROM TELEMARK, SOUTHERN NORWAY

BY

JOHANNES A. DONS

Abstract: Three nodular-rich beds occur within a sandstone unit of the Precambrian Bandak group at Haugli in Telemark, southern Norway. The nodules have ellipsoidal shapes, are 2½—4½ cm long, 1—2 cm across, are composed chiefly of fine-grained quartz, and have a central coarse-grained quartz-filled tube parallel to the long axis. The internal structure of the nodules is insufficiently preserved to permit any definite conclusion as to their origin. An organic or organic-controlled origin (e.g. some primitive sponge or concretionary development influenced by algae) seems more likely than an inorganic origin (e.g. lithophysæ, pisoliths or ooliths).

The name TELEMARKITES ENIGMATICUS has been chosen as a designation for the features.

General introduction to the stratigraphy of the Telemark area.

Geological field investigations in the Precambrian Telemark area of S. Norway are now so far advanced that an essentially complete litho-stratigraphical sequence has been established:

Telemark suite (>4000 m)	*Bandak group*		acid and basic lavas, quartzites, sandstones, tuffs (?), conglomerates, etc.
	— unconformity —		
	Seljord group		quartzites, calcareous sandstones, shales, conglomerates, etc.
	— unconformity —		
	Rjukan group	*Vemork formation*	basic lavas, tuffs, sandstones, etc.
		Tuddal formation	acid lavas with lithophysæ and tuffs, etc. (the "porphyroids")

Rocks belonging to the Telemark suite are found only below the sub-Cambrian peneplain. Because extrusive rocks are typical of the Telemark suite, and the geographically nearby Eocambrian rocks (sparagmites, youngest pre-Cambrian, "proterozoic"), which grade into Cambrian rocks, contain no traces of igneous activity, the Telemark suite and the Eocambrian formations can not be contemporaneous. The Telemark suite is of pre-Eocambrian age. Intrusive rocks of basic composition are abundant, especially as sills, in the Seljord group. The supracrustal rocks have been granitized and intruded by granite locally. Parts of the granites and gneisses surrounding the Telemark suite rocks may represent the original basement upon which the supracrustals were deposited. All the supracrustal rocks are folded.

Absolute age determinations have been performed by the Os/Re method on molybdenite occuring in hydrothermal quartz dikes that cut the Bandak group of rocks (HERR and MERZ, 1958). The results of several determinations give ages of ca. 7×10^8 years. The dikes are young erthan, or contemporaneous with, folding of the Bandak group. The true age of the sediments may be 8-10 hundred million years. A communication on the absolute age determinations of various Norwegian material by several methods will be given in near future by Dr. H. Neumann.

Stratigraphical sequence of the Bandak group at Haugli.

The "fossiliferous", i.e. nodular beds, which are described in this paper were found in the Bandak group in the summer 1957 by the author and his colleague Dr. A. Baer (Neuchâtel, Switzerland) during supplementary field investigations in the "Kviteseid" quadrangle. This area has been mapped for the Geological Survey of Norway by Dr. H. Neumann, and the map soon will be published, together with the adjacent "Rjukan" quadrangle mapped for the Survey by the author (1947—57). A more complete stratigraphical and tectonic description of the Telemark area will accompany these maps. The stratigraphical sequence of the upper part of the Bandak group is listed in the left column of figure 1.

The word "nodule" is used in this paper as a non-genetic term. The letters *a*, *b*, *c*, etc. have no significance except for reference in this paper. They are referred to as str. u. *a* (i.e., stratigraphic unit *a*), str. u. *b* etc.

Bandak group, upper part

h	"quartz sandstone"	> 500 m
g	marble	2–8 m
f	greenstone etc.	> 250 m
e	quartz porphyry	
d	conglomerate	2–10 m
c	greenstone	
b	sandstone	
a	greenstone	

greenstone (str.u. *f*, upper part)	> 200 m
Sandstone with crossbedding	> 30 m
Nodular bed, largest nodules at the top	1,2 m
Sandstone	0,6 m
Alluvium	0,7 m
Nodular bed, small nodules	0,8 m
Alluvium	0,2 m
Greenstone (partly debris)	0,6 m
Nodular bed, large nodules	0,4 m
Nodular bed, small nodules	0,9 m
Porphyry debris	0,5 m
Quartzite	1,0 m
Greenstone aggl. or congl.	1,5 m
greenstone (str. u. *f*, lower part	> 10 m

Fig. 1. Stratigraphical sequence. Upper part of the Bandak group. Left column: the detailed sequence at Haugli. Thicknesses in meters.

Field observations indicate that the sequence of alternating beds, lenses, and tongues of greenstones (str. u. *a*, *c* and *f*, representing basic lavas), quartzites (str. u. *b*, etc.), and a polymictic conglomerate (str. u. *d*) was deposited in a lacustrine environment, probably on the slope of a relatively flat shield volcano.

The rose to grayish quartz porphyry (str. u. *e*) is originally a rhyolitic lava (or welded tuff?). The presence of marble (str. u. *g*) is interesting owing to the extreme rarity of calcareous rocks in the Telemark suite.

Quartz sandstone (str. u. *h*) is a "sack name" for a great variety of rocks, such as quartz schists with some mica (quarried as high-quality whetstone since the eleventh century), shales, and polymictic conglomerates.

Bedrock is exposed in nearly horizontal surfaces approximately 100 square meters in size between the road and the Dalaåi river, at the farm Haugli in the deep valley, Dalane, 10 km NW of Kviteseid (140 km WSW of Oslo). The rocks, undoubtedly sediments (fig. 1, right column), have a distinct primary bedding which now is nearly vertical and strikes generally SSW. The nodular beds were traced for about 300 meters in two directions from Haugli: — to the NE they seem to disappear rapidly; to the S they extend for a greater distance.

The nodules occur in three beds separated by greenstone or sandstone. The greenstone is a mixture of two different irregularly distributed lithologic types. One, a light variety, is undoubtedly a quartz-rich sediment, whereas the other, a deep green type, shows an ophitic texture marked by needles of epidote. Green biotite is abundant. The feldspar is cloudy and, in some sections, nearly absent. The greenstone most probably represents a thin lava flow which incorporated sedimentary material during its flow.

The dark gray sandstone between the upper two and above the upper nodular beds, contains mostly angular quartz, grainsize $< 0,1$ mm and mostly $< 0,05$ mm. The estimated mineral content of a typical specimen is 79 per cent quartz, 10 per cent sericite, 5 per cent calcite, and 5 per cent magnetite (and hematite). Untwinned plagioclase, if present, may have been taken for quartz. In some thin sections traces of sodic plagioclase, alkali feldspar (microcline and perthite) and epidote have been observed. The sandstone between the upper nodular beds shows a well marked stratification with 1—3 mm thick darker layers spaced at ½—2 cm intervals. The dark bands are composed of coarse-grained quartz (diameter 0,15 mm) and laths of secondary alkali feldspar. There is no evidence of graded bedding. Large scale cross-bedding occurs in the sandstone above the upper nodular bed. Although the sandstones are recrystallized, some of the larger quartz grains may reflect original sizes and shapes. Round patches of quartz clusters 1/4—1 cm across, occur in single rows along some of the bedding planes in the vertical beds at Haugli. The patches resemble the nodules described below and are possibly of the same origin. They may, however, be rock fragments.

The contact zone between the sandstone beds and the nodular beds is a few centimeters wide and transitional in character. The border between the greenstone and nodular beds is sharp and parallel to the general stratification.

The nodular beds.

A comparison between the nodular beds shows the following: The lower bed contains small nodules in its basal part (average diameter 9 mm) and larger nodules in its higher part (diameter 20 mm). In the upper bed there is a gradual upward increase in the size of nodules

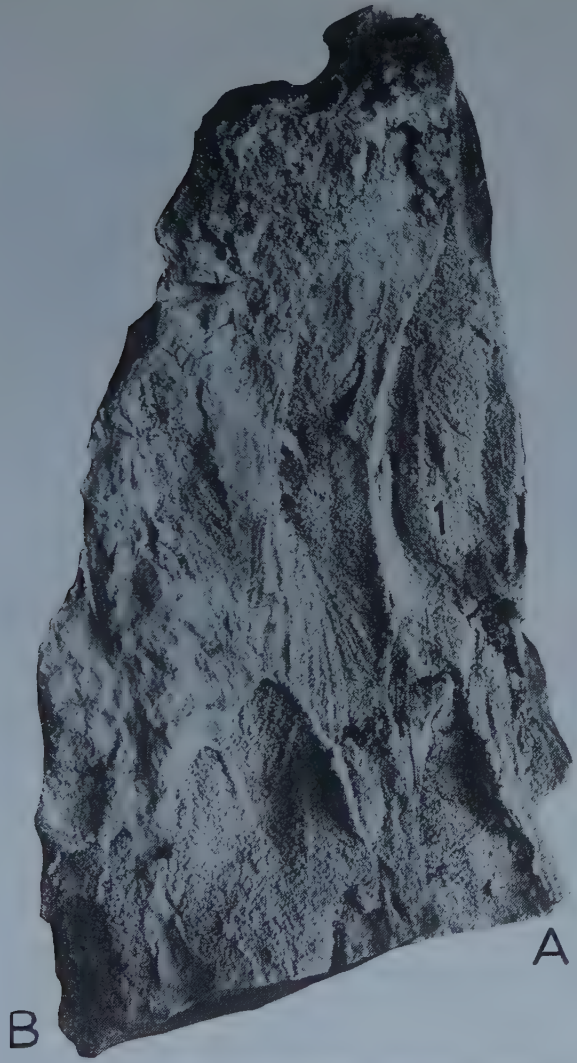

Fig. 2. Specimen from the lowest nodular bed. Greatest length 14 centimeters. Both radial lines (central part) and semicircular pattern (lower middle part) are visible in the depressions. The lump to the right is nodule no. 1 in figures 3, 4, 5, and 6.

from diameter 10 to 20 mm). The diameters were measured on the natural surfaces and give the approximate average size of the short diameter(s) of the nodules. Very small nodules occur in the groundmass.

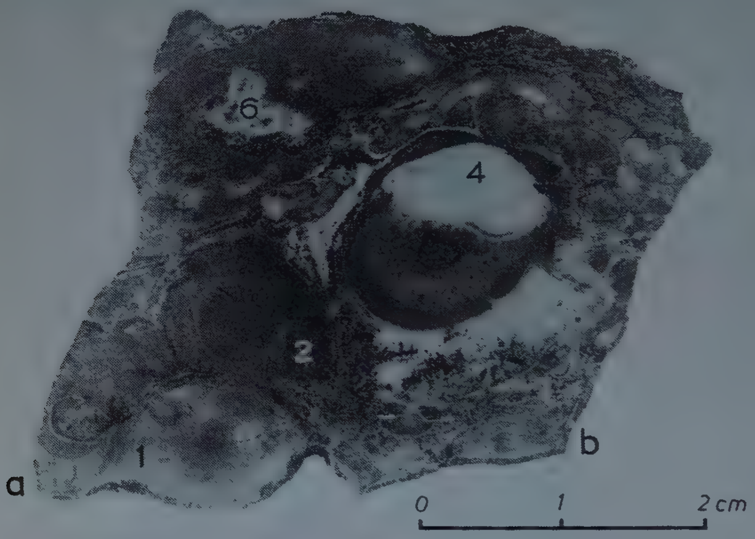

Fig. 3. One of the 21 sections from the specimen of fig. 2. Its location is shown by the a-b line in fig. 6. Nodule no. 4 shows one of the abundant quartz masses described. Nodule no. 2 cut near its end, shows an inner ring and no central tube. To the left of nodule no. 2 and above no. 4 smaller nodules can be seen in the groundmass.

There is no distinct bedding in the nodular rock layers and thus they look unstratified as compared with the sandstones.

The nodular beds are cut by quartz veins of the dilatation type. On some exposed rock surfaces, particularly of the upper nodular bed, the nodules are more or less masked by square to rhomb-shaped aggregates (sidelength 1—3 cm) of quartz and muscovite. These may be pseudomorphs after a mineral which grew in the nodules.

Oriented specimens have been studied by series' of sections. The best results were obtained from a sample taken from the lowest nodular bed (fig. 2). The specimen was cut in 2 mm thick plates normal to the long axis of the nodules (fig. 3). The patterns on the plates, that were made clear by color differences (deep red, green, yellow, and white), were copied by drawing on 2 mm thick perspex sheets. Then, all the perspex sheets were assembled into correct position, (also taking into consideration material removed by the diamond saw) and a transparent reconstruction of the original specimen was obtained. Figures 4 and 5 show two different parts of the perspex model placed on a rubber cast of the original specimen. Fig. 6 shows a drawing based on the perspex model — the nodule outlines were projected to a plane parallel to their long axis and also essentially parallel to the bedding-planes in the sandstones (strike NNE, dip nearly vertical). The nodules are shaped like ellipsoids of revolution, the long axis ranging in length from $2\frac{1}{2}$—$4\frac{1}{2}$ cm and short axis, from 1—2 cm. A perspex reconstruction of one specimen, however, showed relatively narrower cone-shaped ends. On a weathered surface a few nodules broken across their long axis show elliptical sections. All specimens show only one axial tube, parallel to the long axis.

The internal structures of some nodules are partly destroyed by clear, white or greenish quartz masses; which have grown and made the nodules swell and partly burst. The quartz masses occupy completely random positions within the nodules.

Minerals found to occur in both the nodules and the "groundmass" are the following: quartz (main constituent), sericite (partly paragonitic?), calcite, titaniferous hematite, and a few grains of magnetite. The grain size is typically < 0.05 mm for all except the quartz grains in the central tubes of the nodules, in the irregular quartz masses, and in the veins. These latter grains attain up to, 0.6 mm in their greatest dimension. A few grains of oligoclase, microcline, and zoisite occur in the groundmass. Calcite is irregularly distributed, but is most abundant in the groundmass. Treatment with hydrofluoric acid gave no residue thus indicating that carbon is not present. An alkali determination of the whole rock gave the surprising result: 2,04 per cent K_2O and 4,35 per cent Na_2O. The high content of Na_2O may be explained at least to some extent by the possibility that the sericite is paragonitic. Because the sericite content is fairly low, however, it seems more likely that submicrospic or untwinned plagioclase grains do occur in considerable amounts in the groundmass. An alkali determination of material from a nodule gave 0,42 per cent K_2O and 3,88 per cent Na_2O; which shows that probably the nodules also contain plagioclase.

When studied in thin sections (parallel light, low magnification) the nodules are easily seen, because concentrations of hematite and

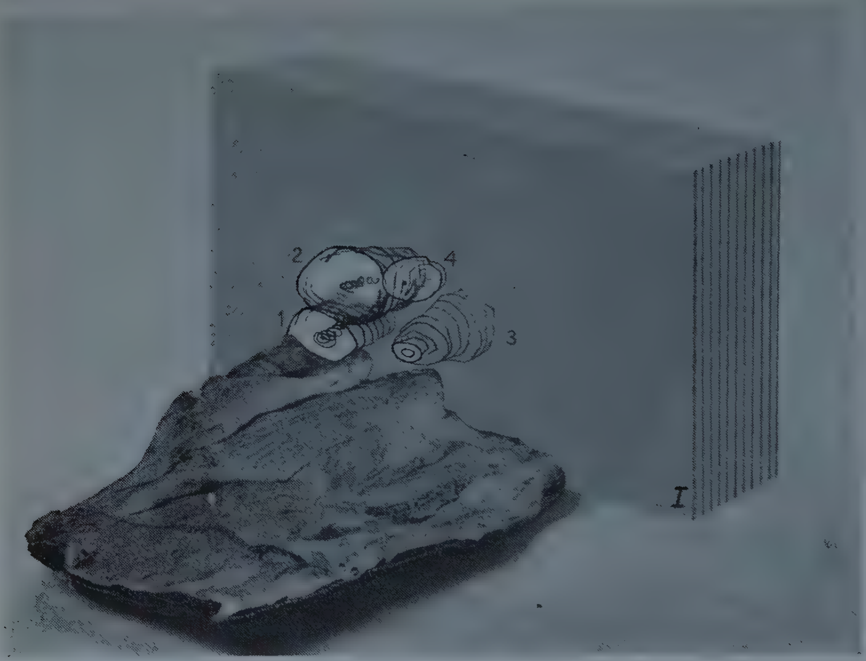

Fig. 4.

Fig. 4 and 5. The lower part of each picture shows a rubbercast (negative) of the specimen of fig. 2. The upper parts consist of perspex sheets that represent the original sectioned specimen. Features seen in the sections have been drawn on each sheet. The outlines and surfaces of the perspex piles have been retouched.

thin layers of pure quartz form alternating gray and white concentric bands near their outlines (fig. 7). The hematite occurs within the quartz grains as well as between them. Between crossed nicols the sections of the nodules show shades of yellowish-green colors, because of local concentrations of sericite. No well defined regular pattern is outlined by these small patches.

Some thoroughly weathered sections of the nodules show both radiating and concentric structures. A section parallel to the axial tube revealed radiating structures branching out from the end of the tube. Sections cut perpendicular to the tubes near the ends of the nodules show concentric rings (fig. 3, nodule 2), indicating termination of the nodules by at least two layers.

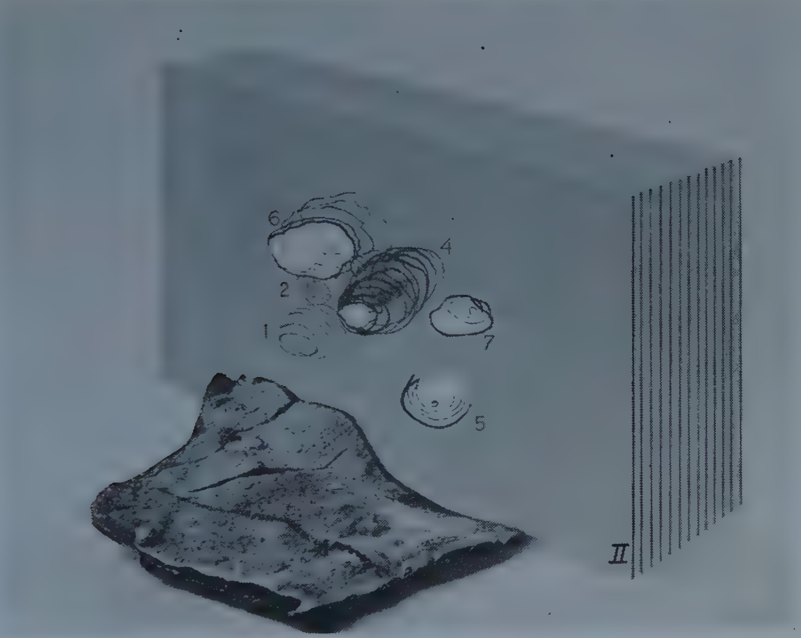

Fig. 5.

From figures 3 and 7 the impression is gained that certain features are common to the groundmass and to the nodules. If so, it seems possible that the radial- and concentric- appearing features in the two depressions of the specimen fig. 2 represent some sort of enclosing material (see fig. 8).

Structurally the Haugli locality is situated on the western limb of a SW-plunging anticline, a large part of which is shown in a general map published by Dr. H. Neumann (Neumann, 1944). The long axis of the nodules is parallel to the bedding planes of the enclosing sediments and plunges ~ 75° SW, apparently corresponding to the poorly defined major fold axis. But 300 m SW of Haugli the long axis is vertical or plunges steeply NNE. There is, however, no geologic evidence to indicate that the anticlinal axis accordingly has reversed its plunge. Moreover, the axial tube of the nodules protrudes farther out at the ends of the nodules than could the central core of a deformed sphere. This evidence, in conjunction with the fact that in the Haugli area the pebbles of the conglomerate, str. u. *d*,are nearly unde-

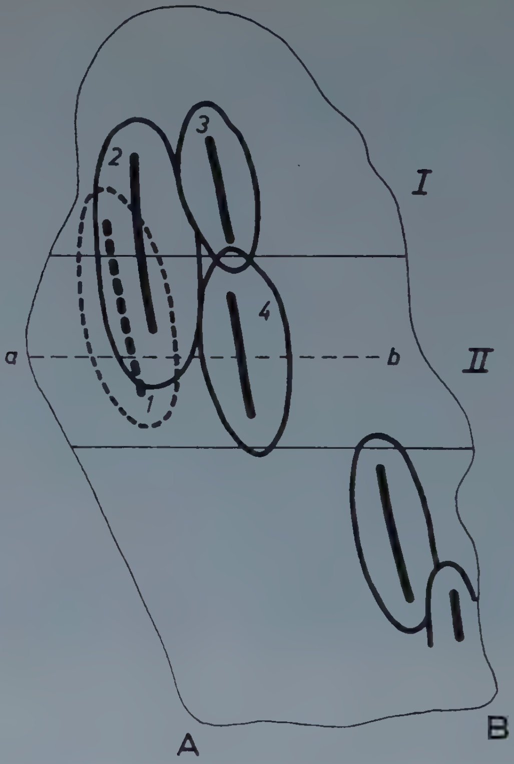

Fig. 6. Most of the nodules shown in fig. 4 and 5 have been projected in this figure into a plane parallel to the bedding of the adjacent well stratified rocks. Letters and numbers correspond to those used in fig. 3, 4, 5.

formed, leads the author to believe that the orientation and shape of the nodules were only slightly effected by the regional deformation which produced the folds.

Discussion.

One main problem remains unsolved: what is the origin of these features? Are they organic, inorganic or both organic and inorganic? Without trying to cover every imaginable possibility some main points will be mentioned below in order to initiate future discussion of the problematica.

No carbon could be identified, and there appears to be no hope of finding amino-acid or other more certain organic compounds in the rock. The present mineralogical composition of the rock (mainly

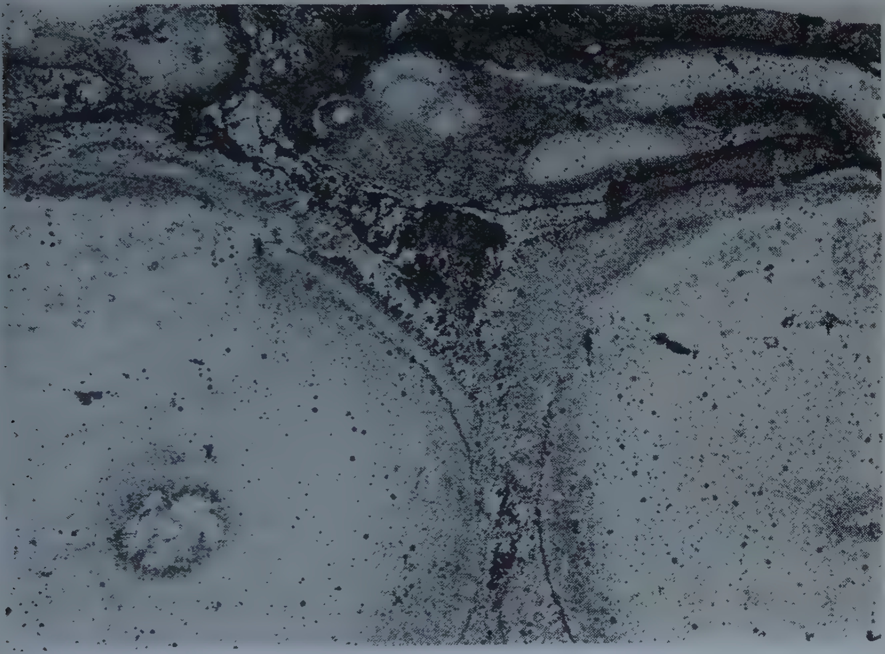

Fig. 7. Photo micrograph, parallel light. The distance between the centers of the nodules is 14 millimeters. (Nodule no. 4 to the left, no. 2 to the right.) Dark spots and lines in the nodules and along their borders are composed of hematite. White patches in the black groundmass may be related to the nodules.

quartz) does not necessarily correspond to the original composition of the sediment which contained the nodules — perhaps, for example, it was a calcareous rock which has been silicified. The rhomb-shaped "pseudomorphs" (now quartz and muscovite) that occur in some of the nodules probably indicate such replacement processes. On the other hand the marble bed (str.u. *g*) that is present somewhat higher within the sequence has not been silicified.

The internal structure of the nodules seems to be insufficiently preserved to permit unreserved acceptance or refusual of the possibility that the nodules are purely or partly organic remains. All the nodules have the same internal structures. There is only one quartz-filled axial tube. Small nodules, probably of the same shape and construction as the larger ones, occur in the groundmass between the large nodules.

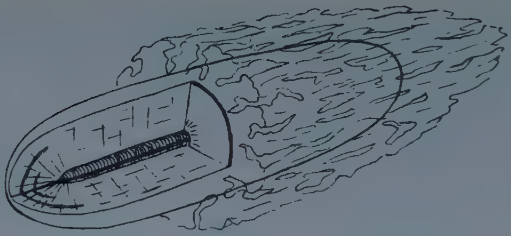

Fig. 8. A simplified reconstruction of TELEMARKITES ENIGMATICUS with the internal structures shown; surrounded by an irregular envelope of doubtful relation.

It is believed that the nodules were formed or deposited in shallow water during unstable conditions including nearby volcanic activity. The "enclosing features" (fig. 8) within the groundmass of the nodules seem to disprove the possibility of conglomeratic origin. If the nodules had been formed elsewhere for instance as lithophysæ or ooliths and subsequently freed by weathering and erosion and transported to their present location, one should expect to find at least a few partly broken nodules. None have been found. In situ formation as lithophysæ is excluded because the nodular beds are of sedimentary origin. The extremely regular and consistent make up shown by these nodules has seldom been seen in oolitic or pisolitic beds. It is believed that the nodules are practically uneffected by tectonic deformation.

There is a striking resemblance between the nodules found in Telemark and a photo reproduced by Prof. A. Hadding representing a "section through the lower part of a stromatoporoid bed showing the centra of growth, Högklint stage" from a Silurian reef limestone of Gotland, Sweden (Hadding, 1941). Hadding writes in a personal letter to the author, after he had seen one photo of a section of the nodules from Telemark: "Your photo undeniably shows a striking resemblance with fig. 11 in my work on Reef Limestones. The stromatoporoid, of which the basal part is figured, has the shape of a dome with diameter 200 mm. Stemlike stromatoporoids exist also, however. The Telemark-fossil I would more likely asume to be a (silicified) calcareous alga". (translation from Swedish).

The material was shown, described, and discussed at the III Winter

Meeting of Nordic Geologists in Finland (III Winter Meeting, 1958). With the hope that other geologists might have seen similar structures, a series of 7 original photo's (most of them reproduced in this paper) was sent along with a short general description to experts in Europe and USA.

The comments may be summarized briefly. Prof. L. Dangeard, Caen, France could see no definite solution of the problem, but considers deformed ooliths or pseudo-ooliths most likely. Dr. E. Flügel, Vienna, Austria, stated that there are no obvious affinities between the oldest known stromatoporoids and the nodules from Telemark. He also mentioned the possibilities of algal-structures or inorganic flattened pebbles. Prof. J. St. Jean, Jr., Chapel Hill, N.C. USA has made several thin sections of the material which was sent him, and he writes: ".... I was also interested in the vermicular "septate" structures in the associated matrix which is suggestive of organic structures, but nothing appears to be identifiable. The specimen, whatever it may be, can not possibly be a stromatoporoid, though it does superficially have the general shape of the beatriceoid stromatoporoid, it has none of the internal structures of that group of fossils ....". Prof. O. H. Schindewolf, Tübingen, Germany also showed the photographs to Dr. Seilacher, Frankfurt a. Main, Germany. Although both men found the occurrence interesting but difficult to explain by seeing only the photos, the former proposed pisoliths formed probably by intervention of algae. Prof. R. R. Shrock, Cambridge, Mass., USA, to whom some material also was sent at request, said in a letter "...... I do not have any strong feeling either for or against an organic origin for the concretionary structures ..... The axial tube certainly does show up clearly and since it is always in the plane of stratification, I would be inclined to think that the structures might represent stems of some kind (provided that we can establish that stems were present at all) that were deposited on the sedimentational interface and then flattened somewhat by the consolidation of the surrounding material". Prof. H. Termier, Paris, thinks that they are concretions formed by the activity of schizophytes, probably cyanophyceae (blue algae).

The writer prefers not to give an opinion as this would not be based on very good positive evidence. After having worked with the material he feels, however, that the future explanation will be one of two: 1) the nodules are concretions formed by intervention of algae, or 2) the nodules represent sponges.

The name TELEMARKITES ENIGMATICUS is proposed for the features. Telemark is a county (fylke) in southern Norway. The type locality is Haugli, Dalane in Telemark. The type specimen is deposited in the type collection of Paleontologisk Museum, University of Oslo with no. 69 586.

The author wishes to thank all persons who have taken interest and given help and suggestions during his investigations of the material.

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Oslo, Norway.

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NOTISER

Om vulkanittene i den kaledonske fjellkjede i Norge.

Foredrag holdt i Norsk Geologisk Forening 12. febr. 1959.

Av

CHRISTOFFER OFTEDAHL

Vulkanske bergarter har blitt beskrevet fra nesten alle deler av den kaledonske sone som går gjennom Norge fra Stavanger til Tromsø, en strekning på omkring 1400 km. Det dreier seg vesentlig om ofiolittiske grønnsteiner og grønnskifere som vanligvis blir antatt å representere metamorfe basaltiske lavaer og tuffer. Dette gjelder særlig Støren—Bymarkgruppen. I denne gruppe finnes det også noen sure vulkanitter beskrevet som slike av Kvale fra Stord (1937), C. W. Carstens fra Løkkenfeltet (1915), Foslie fra Ølve—Varaldsøy i Hardanger (1955) og H. Carstens fra Fosenfeltet (1955).

Foredragsholderen hevdet imidlertid at sure vulkanitter synes å være meget mer utbredt enn det fremgår av litteraturen. Grunnlaget for dette syn er foredragsholderens detaljerte kjennskap til Grongfeltet, kartlagt med stor nøyaktighet av Foslie. I grønnsteinsformasjonen er det her meget vanlig å finne lyse benker av såkalt kvarts-keratofyr som er meget utholdende etter lagflatene. Man må anta at det helt vesentlige er pyroklastiske sedimenter av følgende grunner: Der er skarp grense eller gradvis overgang til grønne tuffskifere, der er alle slags vekslinger mellom lyse keratofyriske lag og mørke basiske tufflag, fra alternerende, metertykke benker ned til cm- og mm- tykke skarpe bånd; dertil er kvarts-keratofyrer også i noen tilfeller assosiert med grove agglomerater. Videre er granittisk magma så seigt at det nødig danner langtrekkende lavaer; ennu vanskeligere danner det utstrakte, flattliggende "sills" uten helt spesielle geologiske grunner (eks.: Mænaittganger på grensen mellom prekambrium og kambrosilur i Oslofeltet).

Når alternerende lag av kvarts-keratofyr og grønnsteinstuff blir utsatt for sterkere regionalmetamorfose, omkrystalliserer de til henholdsvis middelkornig massiv trondhjemitt og amfibolitt. Vekslende lag av slike bergarter synes å være alminnelige i fjellkjeden, og foredragsholderen hevdet at det vesentligste av slike bergarter, ofte beskrevet under betegnelsen «injeksjonsgneis», sannsynligvis representerer metamorfoserte pyroklastiske sedimenter.

Foruten kvarts-keratofyr og «injeksjonsgneis» er det beskrevet alle overgangsledd fra kvarts-keratofyr via trondhjemitt til granitt, i form av lag eller «sills», antatt å være intrusjoner. Alle disse bergarter bør mistenkes for å være opprinnelige suprakrustaler. De finnes i følgende områder:

Stavangerfeltet etter beskrivelser av Goldschmidt (1921), Osdistriktet etter beskrivelser av Reusch (1882), N.-H. Kolderup (1935) og C. F. Kolderup og N.-H. Kolderup (1940), videre Norheimsundområdet etter beskrivelse av Kvale (1945), muligens Sunnfjordområdet ifølge kort beskrivelse av N.-H. Kolderup (1955); i Trondheimsfeltet er båndete bergarter kjent fra Ottaområdet ifølge Strand (1951), keratofyrer fra Rørosområdet ifølge Carstens (1910) og fra Hølonda—Horg området etter Th. Vogt (1945). Keratofyrbenker er kjent fra Fosenfeltet ifølge H. Carstens (1955), fra Grongfeltet ifølge Oftedahl (1958). Videre er tynne utholdende benker av granittisk til trondhjemittisk sammensetning beskrevet fra Sulitjelmafeltet av Th. Vogt (1928) og fra kartbladet Tysfjord ved S. Foslie (1941). Foslie bringer også fotografier av de typiske skarpt alternerende lag av keratofyr eller finkornig granitt og hornblendeskifer. Konklusjonen er således at den underordovisiske vulkanisme var meget blandet, med utstrømninger av submarine basalter og med dannelse av pyroklastiske produkter i store mengder fra basisk til sur sammensetning.

Ifølge den forhåndenværende litteratur er de sure eller granittiske lag dels av vanlig granittisk sammensetning, men oftest av trondhjemittisk sammensetning (ned til 1 % K_2O) og keratofyrisk sammensetning (5—7 % Na_2O og ned til 0,1 % K_2O). Denne siste sammensetning dominerer muligens i de jevne og utholdende benker som kan tenkes å representere langt transporterte asker. Foredragsholderen hevdet at denne keratofyriske sammensetning umulig kan være primærmagmatisk og foreslo at den kan være et resultat av en havvannets natronmetasomatose overfor de fine askepartikler.

I en rekke geologiske beskrivelser fra fjellkjeden har det vært omtalt finkornige hornblendegabbroer som små flekker og linser eller større fakolitter i grønnstein eller glimmerskifer. Til dels har det vært observert gradvis overgang fra grønnstein til hornblendegabbro (av Kvale på Stord, 1937, og av Th. Vogt i Sulitjelmafeltet, 1928). Foredragsholderen har sett at dette er et meget vanlig fenomen i Grongfeltets grønnsteiner og oppfatter disse gabbroer som effusiver som har krystallisert til grovere tekstur under et visst lite dekke av de tilsvarende lavaer. Dette er i overensstemmelse med et forslag av Vogt fra 1928, og det er den stadig mer utbredte oppfatning av slike ofiolittiske gabbroers dannelse omkring Middelhavet (f.eks. Dubertret i Syria, 1952).

Foredragsholderen omtalte videre nyere hypoteser angående dannelsen av grønnsteiner og serpentiner, særlig av Rittmann, 1958. Denne oppfatter grønnsteinsdannelsen som primær, d.v.s. at basaltlavaen har krystallisert under stort vanntrykk mer eller mindre direkte til grønnstein. Han oppfatter også serpentinene som olivinrik lava som under stort havtrykk har krystallisert direkte til serpentin. Et feltbevis på dette har Rittmann funnet i Eastern Desert i Egypten, hvor serpentin med putestruktur ligger mellom umetamorfe sedimenter. Denne antagelse synes med god grunn å kunne gjøres gjeldende for serpentinene i Kaledonidene, idet serpentinene ofte definerer et ganske bestemt nivå, av mange oppfattet som et bestemt intrusjonsnivå. Det er imidlertid mer plausibelt å oppfatte dette som forårsaket av en bestemt ekstrusjonsfase.

Bemerkninger til dr. philos. Christoffer Oftedahls foredrag i Norsk Geologisk Forening, 12. februar 1959.

I sitt meget inspirerende foredrag berørte dr. Oftedahl en serie av de problemer en fjellkjedegeolog stadig støter på og stadig forundrer seg over.

Det har alltid vært litt av et tankekors å forstå hvorledes utholdende, tynne lag av granittisk sammensetning kan være oppstått ved injeksjonsmetamorfose. Tidligere har man da også grepet til metamorf differensiasjon som noe av et universalmiddel. Vi kjenner de geologer som så å si ikke tror på magma, men forklarer alt mulig ut fra metamorf differensiasjon.

Når nu dr. Oftedahl setter inn sure tuffer som et universalmiddel og dermed mener å ha løst hele «fjellgåten», så er det mulig at han har grepet endel av sannheten og det ansees til og med for sannsynlig at han i mange tilfelle virkelig har funnet løsningen på problemer som har vært vanskelige å forklare. Det forekommer imidlertid som om dr. Oftedahl har sprunget vel lett over endel av de vanskeligheter man har med å gjøre i denne forbindelse. De asker som har bygget opp tuffene må i det vesentlige forutsettes å ha hatt en glasskarakter og noe nær granittisk kjemi. Når slike glass utsettes for forvitring i vanlig miljø, vil de i alminnelighet gå over til å danne enten bentonitter eller hydroglimmermineraler, kfr. Erik Norins påvisning av nydannet illit i Det Tyrrhenske Hav (1958). Det er vanskelig å se at sedimentene av slike glass kan ha undergått en utpreget natriummetasomatose i det vanlige miljø på bunnen av havet. De kjemiske prosesser som eventuelt her er skjedd vil antagelig ha ført til dannelse av kalium- og kalsiumrike bentonitter, slik som bentonitthorisontene i Oslo-feltet og Kinekulle, kfr. Anne Marie Bystrøm (1954), Fredrik Hagemann og N. Spjeldnæs (1955). Dannelse av natriumbentonitter vil ikke forekomme. På tross av at havvannet bare holder 0,38 gram kalium pr. liter og 10,55 gram natrium pr. liter, er den primære virkning på glassmineralene at det oppstår en kalimetasomatose og ikke en natrium-metasomatose. Det er derfor ikke lett å forstå når Oftedahl hevder at sjøvannet på havbunnen har foretatt de merkelige reaksjoner som skal til for å få dannet så ekstreme bergarter som han forteller om, med natriuminnhold av størrelsesorden 7 % og et kaliuminnhold fra 0,2 og helt ned til 0,02 %. Dette er åpenbart en sammensetning som må være forskjellig fra den de hypotetiske asker engang kan ha hatt. Det må ha skjedd en kjemisk separasjon som i sin effektivitet ligger over den effektivetet en vanlig god analytiker har i sitt kjemiske laboratorium. Denne separasjon kan vanskelig ha foregått i havvannet mens glasskornene svevet omkring eller på havbunnen, der i et uherdet slamsediment. Hertil kommer at den har vært utsatt for metamorfose som oftest svarer til høyere deler av grønnskiferfasis, og dette forutsetter en viss temperatur og et betydelig vanndamptrykk. Dr. Oftedahl forutsetter at dette trykk skal være hydrostatisk på grunn av ovenforliggende vannsøyle. Dette krever imidlertid vanndyp som må anslås til størrelsesorden 20—50 km, kfr. Fyfe, Turner og Verhoogen (1958), som antar dyp i jordskorpen på 10—30 km i våte sedimenter. Selv om man bare bruker den laveste av disse verdier og regner om til vannsøyle, kommer man til tall som ikke på noen måte kan bringes i overensstemmelse med de gruntvannsavsetninger (kalkhorisonter) man finner enkelte steder som et av de sterkeste indisier på at man har med suprakrustale bergarter å gjøre. Derfor vil jeg både av kjemiske og av mekaniske grunner anse det for usannsynlig at de prosesser dr. Oftedahl forutsetter har funnet sted på den måte han antar, men at det er mulig og ikke strider hverken mot faselæren eller kjente kjemiske prinsipper at tilsvarende reaksjoner til de dr. Oftedahl antyder kan finne sted på litt større dyp. Dersom et sjøvannsmettet, meget porøst sediment, eventuelt tuff, overlagres av et impermeabelt eller lite permeabelt lag av en basisk lava eller leire og derefter med sitt totale sjøvann bringes ned til et dyp der temperaturen går opp i ca. 300°, kan det tenkes å foregå en omsetning, hvorved det avdestilleres en kaliumkloridinnholdig flyktig fase som eventuelt er istand til å besørge kali-metasomatose i høyere lag. Det residium man har igjen blir da en utpreget natriumbetonet bergart svarende til de ekstreme bergarter som dr. Oftedahl referer til.

Ivan Th. Rosenqvist

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Svar til Ivan Th. Rosenqvist.

Tiden tillot ikke nærmere diskusjon av spilittiseringen av sure askelag i mitt foredrag. Til prof. Rosenqvist's innvendinger mot å oppfatte kvartskeratofyrenes natronovervekt som et resultat av reaksjon mellom havvannet og glødende, surt askeregn kan jeg svare følgende.

Prof. Rosenqvist anser det mere sannsynlig at det høye natroninnhold og det meget lave kali-innhold stammer fra regionalmetamorfose-reaksjon mellom det vannrike aske-sediment og det natrium som fantes i porevannet. Denne mulighet har vært fremhevet av mange, f.eks. av R. Daly i 1914, G. Beskow i 1929 og C. F. Park i 1946 (se litteraturreferanser hos Amstutz, 1958). Jeg er enig i at det er en god mulighet, men det er en rekke trekk som gjør at jeg holder den i foredraget nevnte prosess for å være den sannsynligste.

Høyt natroninnhold og usedvanlig lavt kali-innhold (omkring 0,1 %) er vanlig ikke bare i kvartskeratofyrer, men også i basiske tuffskifre. Det samme gjenfinnes i basaltiske putelavaer, se Th. Vogt (1945, p. 466—468). Her ligger oftest K_2O-innholdet på 0,1—0,2 % mens platåbasaltene har 0.7—0,8 %. Den merkelige K-fattigdom ser jeg nødvendigvis som et resultat av spilittiseringen, og angående denne støtter jeg Rittmanns oppfatning (1958): Spilittiseringen av submarine basalter (d.v.s. nedbrytning av plagioklas til epidot og albitt, dannelse av hornblende og kloritt istedenfor pyroksen og fjernelse av det meste kalium eller erstatning med natrium) skjer ved basaltenes utstrømning på havbunnen ved reaksjoner med havvann under betydelig trykk. Den helt kontinuerlige overgang til sure, Na-rike askelag gjør at jeg må anta en slik nesten instantan reaksjon også for de sure askers alkaliforhold.

En prosess som synes å være beslektet med den her foreslåtte prosess er bentonitt-dannelsen av sure askelag. I følge Grim (1953, s. 363) har de fleste bentonitter Ca som det viktige utbyttbare jon, men det finnes også bentonitter som er særlig natronrike og kalifattige, —

og omvendt. De geologiske forhold tyder meget sterkt på at omdannelsen av asken til bentonitt fant sted like etter avsetningen eller muligens nesten samtidig med avsetningen. «Bentonittiseringen» må altså ha funnet sted som en reaksjon mellom aske og havvann, ofte på grundt vann, og reaksjonen er en devitrifisering av ganske bestemt type, ledsaget av enten utlutning av alkalier (det vanligste) eller av K- eller Na-metasomatose (sjeldnere). Selve prosessene er lite forstått, men det syns ikke urimelig å foreslå en lignende Na-metasomatose for de askeregn som nu fremtrer som orogene kvartskeratofyrer.

Chr. Oftedahl

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Note on the equilibrium between plagioclase and epidote

By

OLAV H. J. CHRISTIE.

New data concerning the exsolution of plagioclases appear to modify the equilibrium curve between plagioclase and epidote.

P. ESKOLA (1914) and J. H. L. VOGT (1927) contended that if sufficient water is present for the formation of epidote, plagioclase and epidote are unstable together. T. F. W. BARTH (1928 and 1936), T. STRAND (1942), I. ROSENQVIST (1942) and J. A. W. BUGGE (1943) have discussed the equilibrium between plagioclase and epidote, and H. RAMBERG (1944) proposed a tentative diagram for the stability relations between these two minerals. Ramberg's diagram was based on empirical data, and because of experimental difficulties the system has not been experimentally examined.

According to the diagram (Fig. 1) a plagioclase of composition X_2 would be stable under moderate water pressure at temperature T_1. At temperature T_2 epidote will form and take some CaO, Al_2O_3 and SiO_2 from the plagioclase. The plagioclase therefore will be enriched in Na_2O. The minerals in equilibrium at T_2 would be epidote of composition X_1 and plagioclase of composition X_3 provided sufficient

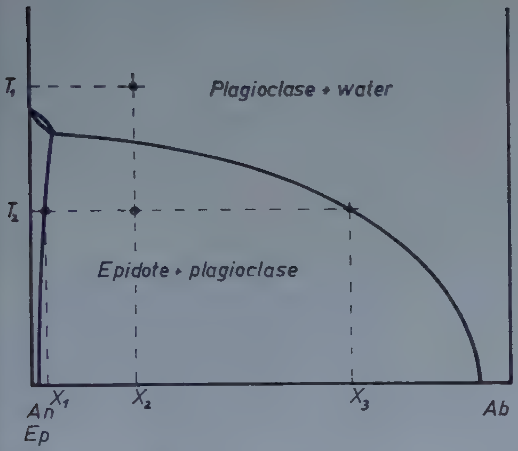

Fig. 1. Equilibrium curve between plagioclase and epidote slightly modified after Ramberg (1943). This diagram is an approximation, because the reaction plagioclase + water = epidote is stoichiometrically not quite correct.

water is present for the formation of epidote. Thus the An content of the plagioclase in equilibrium with epidote under moderate water pressure would decrease with decreasing temperature.

Plagioclases consisting of lamellae of different composition were called peristerites by O. B. Bøggild (1924). S. Chao and W. H. Taylor (1940) suggested that some intermediate low temperature plagioclases consist og lamellae of An rich and An poor plagioclases and thus are peristerites. W. F. Cole, H. Sørum and W. H. Taylor (1951) studied the structure of plagioclases and showed that low temperature plagioclases ranging from An_{30} to An_{72} in bulk composition form mechanical mixtures of An rich and An poor components.

Careful examinations by F. Laves (1954) indicate another low temperature exsolution field between An_5 and An_{17}. L. Gay and J. L. Smith (1955) confirmed this and extend the exsolution field to An_5 — An_{23}

Thus there would seem to be two exsolution fields in the sub solidus area of the plagioclases. A plagioclase of composition Y_0 (Fig. 2) will split up to form two plagioclases of composition Y_1 and Y_2 respectively, when heated under dry conditions for a long time at the temperature T_3. Because the separation of a given plagioclase into two plagioclase phases of different Na/Ca ration involves diffusion not only of Na and Ca, but also of Al and Si, the formation of peristerites is very slow at lower temperatures. (J. Goldsmith, 1952). Below a certain temperature diffusion of Al and Si within the plagioclase would be so slow that the formation of peristerites is negligible even in geological time. Nevertheless, plagioclases within the exsolution field are unstable phases exhibiting higher free energy than do the stable phases (peristerites) and therefore possessing higher chemical reactivity. If one of the peristerite members reacts to form new minerals, the exsolution probably would be faciliated.

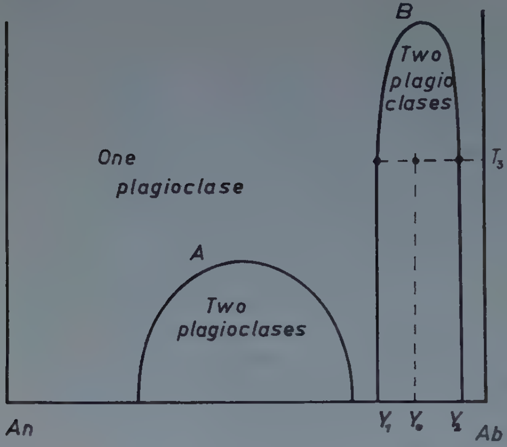

Fig. 2. Exsolution fields at subsolidus temperatures for plagioclases. Curve A after Cole, Sørum and Taylor (1951) curve B after Gay and Smith (1955). Temperature scale unknown.

The phase diagram fig. 3 is based upon these considerations. At moderate water pressure and a given temperature T_5 a plagioclase of composition Z_0 should split into epidote of composition Z_1 and plagioclase of composition Z_2. This plagioclase (Z_2) is unstable and splits up into Z_2 and Z_4 plagioclase, following curve B of fig. 2. However, under the present conditions, the plagioclase Z_3 is not stable in contact with epidote but breaks up into Z_1 epidote and Z_4 plagioclase. Not until the composition Z_4 is reached, will the plagioclase become stable. Thus the final product is Z_1 epidote Z_4 plagioclase. The equilibrium curve between plagioclase and epidote is therefore the full line ABC of figure 3.

Epidote in contact with plagioclase containing between 5 and 23 per cent An would seem to represent unstable phases (composition

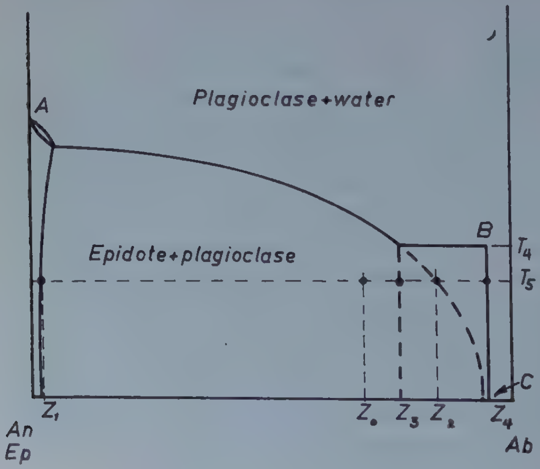

Fig. 3. Equilibrium curve between plagioclase and epidote. The exsolution area corresponding to B in fig. 2 is between Z_3 and Z_4.

between Z_3 and Z_4, temperature below T_4 of figure 3). Therefore, with our present knowledge of the subsolidus relations of the plagioclases, we can not use the equilibrium between plagioclase and epidote as a general geolgical thermometer.

Mineralogisk-geologisk museum, Tøyen, May 1959.

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Crystallization experiments with alkali olivine basaltic glass from Egersund.

By

OLAV H. J. CHRISTIE

Some dikes of alkali olivine basaltic dolerite from the Egersund area contain glass as a border zone. This glass was heated partly dry, partly under controlled water pressure. The dry heating was made in nichrome or platinum furnaces, the heating under water pressure was made either in cold seal bombs or in a bilaterally heated hydrothermal bomb. Similar experiments have been made by H. S. YODER jr. and C. E. TILLEY (1956).

Table 1

Weight per cent		Mol norm		Weight per cent		Weight norm	
SiO_2	45,05	Q	4,97	SiO_2	49,16	Or	2,78
TiO_2	2,83	Or	17,02	TiO_2	2,29	Ab	17,82
Al_2O_3	16,93	Ab	12,67	Al_2O_3	13,33	An	25,30
Fe_2O_3	1,45	An	18,50	Fe_2O_3	1,31		
FeO	11,63	C	6,38	FeO	9,71	Di	22,93
MnO	0,16			MnO	0,16	Hy	15,35
MgO	4,91	En	14,46	MgO	10,41	Ol	9,14
CaO	7,26	Fs	13,87	CaO	10,93		
Na_2O	1,33			Na_2O	2,15	Mt	2,09
K_2O	2,71	Mt	1,60	K_2O	0,51	Il	4,41
P_2O_5	2,82	Il	4,17	P_2O_5	0,16	Ap	0,34
H_2O+	3,51	Ap	6,34	H_2O+	0,04		
H_2O-	0,13			H_2O-	0,05		

Composition of 1. studied glass, anal. Christie, and 2. of glass used by YODER and TILLEY (1956).

Plagioclase formed in the glass in all the runs, and magnetite or hematite was mostly present. In some of the runs pyroxene formed a as stable phase. In some of the runs quartz formed metastably.

Petrological examination of the glassy selvages showed that plagio-

Table 2.

°C	Water pressure bars	Time	Product
650	heated in air	3 months	He + Plag + Q
850	—	1 week	He + Plag
1175	—	1 hour	He + Plag
275	790	3 months	Mag + Plag
280	410	3 months	Mag + Plag
580	1120	1 week	He + Plag + Q
600	1560	2 days	He + Plag + Q
630	450	1 day	He + Plag + tr. Pyr.
800	1380	10 minutes	Plag + Pyr + Q
1000	50	5 minutes	He + Plag + Q
1000	600	5 minutes	Melt.

Selected data from crystallization experiments with alkali olivine basaltic glass from Egersund. He: Hematite Mag: Magnetite Plag: Plagioclase Pyr: Pyroxene Q: Quartz.

clase and pyroxene had formed simultaneously under the pressure and temperature conditions existing during the intrusion of the rock. From table 2 and figure 1 it becomes apparent that the pressure-temperature conditions during the intrusion of the dikes correspond to the dotted area within figure 1.

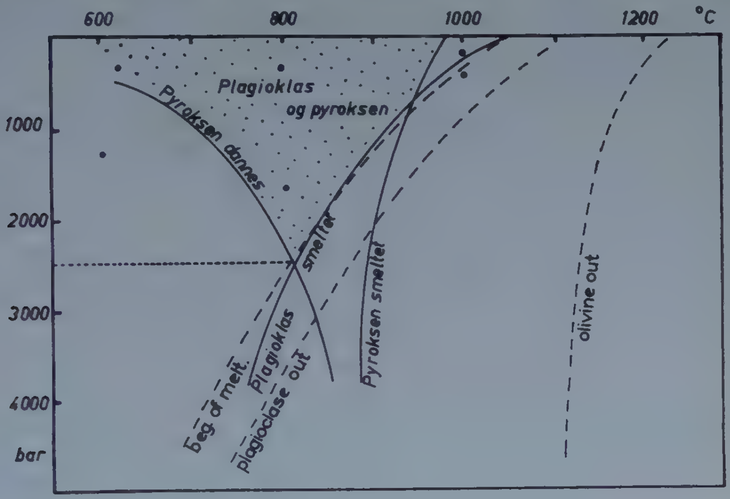

Fig. 1. Full lines, norwegian text: curves constructed from the data of table 2. Dashed lines, english text: curves from Yoder and Tilley (1956).

Thus the temperature of the intruding melt would seem to be between 1000° and 700° C, and the water pressure to be less than 2500 bars.

Mineralogisk-geologisk museum, Tøyen, June 1959.

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H. S. Yoder jr. and C. S. Tilley: Ann. Rept. of the Dir. Geoph. Lab. Wash. 1955—56, p. 169.

Nordisk samarbeid innen geokjemisk prospektering.

Det har lenge vært et ønske om nærmere kontakt mellom institusjoner og personer i de nordiske land som arbeider innen området geokjemisk prospektering. Et skritt i retning av mer samarbeid var et møte som ble holdt i Geologiska Forskningsanstalten, Otaniemi, Finnland i tiden 27.—28. april 1959.

Initiativtaker til dette møtet var geolog L. K. Kauranne, Helsinki. Møtet var først tenkt som en helt uformell sammenkomst av mer privat karakter. Sammenkomsten artet seg imidlertid mer som en kongress hvor i alt 14 personer deltok. Av disse var 1 fra Danmark, 8 fra Finnland, 3 fra Sverige og 2 fra Norge.

Det ble holdt foredrag av herrene Dahlman, Fortescue, Hyvärinen, Kauranne, Landergren, Lupander, Nuutilainen, Salmi, Wennervirta og Viluksela. Videre diskuterte man det fremtidige samarbeid. Som aktuelle samarbeidsformer ble nevnt: Arrangering av møtevirksomhet, utbytting av metoder, resultater og forskere, utdeling av litteratur, standardisering av arbeidsmåter og karter, kooperasjon i større undersøkelser og forskningsprosjekter. Man var enige om at det for å formidle et slikt samarbeid trenges et organ. En komité ble derfor nedsatt. Denne komite skal være midlertidig og ha som hovedoppgave å forberede et nytt møte neste år. Komiteén skal således fungere inntil dette møte velger den endelige komité, eller på andre måter gjør vedtak om fremtidige samarbeidsformer. Det ble besluttet at det neste møte skal søkes lagt til Uppsala i tilknytting det geologiske vintermøte i januar 1960.

Som medlemmer av den midlertidige komité ble valgt: K. Ellitsgaard-Rasmussen, Danmark, L. K. Kauranne, Finnland, B. Bølviken, Norge og S. Landergren, Sverige.

Hvert medlem skal etter forutsetningene inntil videre være en kontaktmann i sitt eget land. Interesserte i Norge kan derfor henvende seg til: *Bjørn Bølviken, adresse Statens råstofflaboratorium, Trondheim,* der man også kan få fyldigere referater fra kongressen.

B. B.

IV Nordiska geologiska vintermötet

vil bli holdt i Uppsala 7.—9. januar 1960. Nærmere opplysninger om møtet fås ved henvendelse til Sekretariatet för IV Nordiska geologiska vintermötet, Mineralogisk-geologiska institutionen, Kåbovägen 2—4, Uppsala.

LOW TEMPERATURE ALKALI FELDSPAR SERIES

BY

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Abstract. Mixtures of microcline and low albite containing different proportions of the two feldspars have been heated at 1025° C and the shift of $d_{20\bar{1}}$ lines has been used to evaluate the extent of homogenisation. It has been assumed that during the short period of heating (96 hours) no appreciable disorder of Al-Si sets in, and with this assumption, a tentative subsolidus diagram for the microcline — low albite system has been proposed.

Natural orthoclase perthites have been heated at 1025° C for different intervals of time and the extent of homogenisation has been studied as a function of time using 131 doublet separation value as an index. It has been found that no homogenisation takes place for the first two hours, after which rate of homogenisation becomes rapid, and after 8 hours the homogenisation is practically complete.

Occurrence of perthites in many rocks and, in some cases, their homogenisation after short heating led one to conclude that soda and potash feldspars form a solid solution at higher temperature which break up into two components on lowering of temperature. On this basis many schematic diagrams have been proposed to illustrate the phase relations.

It is now recognized that there are high- and low temperature modifications of alkali feldspar. In the crystalline lattice the distribution of Si and Al is ordered in the low-temperature modifications, but disordered in the high-temperature modifications.

In 1950, BOWEN and TUTTLE demonstrated the existence of a complete solid solution at high temperatures between the disordered modifications of K and Na feldspar, and by determining the minimum temperature at which unmixing starts for different composition, they gave the solvus for these feldspars (sanidine and high albite), using the variation of the position of $d_{20\bar{1}}$-line in the X-ray powder diagram as a measure of the composition.

But the ordered, low-temperature modifications of alkali feldspar likewise form solid solutions. LAVES (1952) states: "There is thus a strong indication that above certain temperatures there exists (at least pseudostably) a series of solid solutions between microcline and albite. At lower temperatures, however, the alkali feldspars show but very little solid solubility at either end of the series. Therefore an exsolution curve similar to the one determined by BOWEN and TUTTLE (1950) for the sanidine-analbite series must also exist for the microcline-albite series."

As far as experimental work was concerned, he however observed: "The only way at hand to obtain the microcline albite exsolution curve is to determine the lowest temperature of homogenisation of natural microcline-albite perthites (or to determine the unmixing of such a homogenous series). There seems to be little hope of using synthetic material, which is observed to belong always to the high temperature series. Unfortunately suitable material was not at hand for experimental determination." In a recent paper ST. HAFNER and LAVES (1957) mentioned the formation of mixed crystals of albite and microcline by heating albite-microcline mixtures under high pressure for 48 hours.

The above is not a serious difficulty. We can get homogenous microclines with less than 1% of albite in solid solution and albites with similar small amounts of potash feldspar. Chemical analysis can fix the amounts of soda and potash in each phase. To prepare synthetic mixtures, we can weigh such pre-calculated quantities of the above natural materials as would make the mixture of desired composition. For this purpose I selected a microcline from Støa (lent by professor Barth) and an albite from Rømteland, S. Norway (lent by Mr. Sverdrup), both of which are low temperature forms and are found to contain less than 1% of the other constituent in them. Potash feldspar and albite are ground separately to a very fine state and mixed in proper proportions so that the desired compositions are obtained. The mixtures have been briquetted in a hand press and heated at 1025° C ± 3 for 96 hours to effect homogenisation. The partial homogenisation at this temperature (complete homogenisation in three cases) occurring in the mixtures will be evident from the following discussion.

The samples were heated at 1025° C ± 3 for 96 hours. In order to determine the degree of disorder introduced in the Al-Si distribution a sample of the pure microcline was heated. The position of the $20\bar{1}$ peak and the triclinity value of pure microcline is not changed by 96 hours of heating. Thus microcline retained almost its original degree of order after 96 hours of heating and the above thermal treatment alone cannot cause a change in the position of $20\bar{1}$ peak. BOWEN and TUTTLE (1950) and SPENCER (1930, 1937), on the basis of optical and other studies with microcline microperthites, have also concluded that the feldspar retains its low temperature form and character if heated for a short period even at high temperature, and only prolonged heating "at temperatures but little below the melting temperatures" can convert into high form (BOWEN and TUTTLE 1950, p. 502). GOLDSMITH and LAVES (1954) also noticed that to effect Al-Si disorder (i.e. convert the low temperature feldspar to high temperature forms) in dry condition, the feldspars have to be heated for long periods. The position of the $20\bar{1}$ peak of the microcline, as also that of albite, however, change when the mixture of albite and microcline are heated at 1025° C ± 3 for 96 hours (vide table 1); $d_{20\bar{1}}$ decreases in the K-feldspar phase when heated with larger and larger proportions of albite, indicating perhaps a progressive contraction of the unit cell as more and more atoms of K in the unit cell are replaced by Na. Again $d_{20\bar{1}}$ increases slightly in the Na-feldspar phase when heated with increasing proportions of the K-feldspar, indicating perhaps an expansion of the unit cell due to replacement of Na atoms by K atoms. To study the degree of homogenisation from the movement of $20\bar{1}$ peaks, samples are taken from pure microcline, microcline containing 10%, 20%, 30%, 40%, 50%, 60%, 70%, 80%, 90% albite, and pure albite. These are heated for 96 hours at 1025° C ±3, taken out of the furnace and cooled at room temperature (20° C)after which X-ray powder patterns of each of the samples are made. Table 1 shows the $d_{20\bar{1}}$-values of the low albite and microcline phases in the synthetic mixtures of the above different compositions.[1]

---

[1] It is assumed that also in albite very little disorder will be introduced by such thermal treatment as in the case of microcline.

Table 1. Position of $20\bar{1}$ peaks after heating for 96 hours at 1025° C of various mixtures of microcline and albite.

Composition of alkali feldspar mixture K-Fel : Na-Fel	Potash Feldspar phase $20\bar{1}$ peak in A° units	Soda Feldspar phase $20\bar{1}$ peak in A° units	Remarks
100 : 0	4.221	absent	
90 : 10	4.174	absent	Homogenisation
80 : 20	4.169	absent	do.
70 : 30	4.166	Faint peak	Partial homogenisation (?)
60 : 40	4.159	4.067	do.
50 : 50	4.151	4.057	do.
40 : 60	4.137	4.050	do.
30 : 70	4.131	4.041	do.
20 : 80	4.131	4.041	do.
10 : 90	absent	4.041	Homogenisation
0 : 100	absent	4.037	

(The X-ray powder patterns are taken by a Nonius Guinier Camera of effective diameter of 22.98 cms).

On the basis of the homogenisation experiments done by me (vide table 1), I intend to discuss certain aspects of the miscibility curve of microcline-albite system. It has already been argued that though the homogenisation is done at 1025° C, the microline remains as microcline and low albite as low albite, and only K and Na ions migrate from one phase to the other causing complete or partial homogenisation.

It will be seen from table 1 that for mixtures Ab 10% Mi 90% and Ab 20% Mi 80%, one of the phase disappears after heating, and the $d_{20\bar{1}}$ line for one of the phases is absent; in other words complete solid solution occurs in these ranges. So is also the case for mixture of composition Ab 90% Mi 10%. Between compositions Ab 30% Mi 70% to Ab 90% Mi 10%, however, we get two separate coexisting phases, one potash rich and another soda rich, having different $d_{20\bar{1}}$ values.

It appears to me that the low temperature potash phase can take less than 30% of albite in solid solution and beyond this composition two phases appear. The two phases do not tend to homogenise further by heating at 1025° C for a period much longer than 96 hours, as indicated by the fact that $d_{20\bar{1}}$ values for two phases do not change appreciably any further. Probably the following important inference can be drawn from this fact. When the mixture under investigation is heated at 1025° C for 96 hours an equilibrium with respect to Na : K ratios in these phases corresponding to a certain temperature (much lower than 1000° C) is reached, but a reaction barrier prevents the reaching of equilibrium expected at 1025° C; the barrier is nothing but the order-disorder potential barrier with respect to Al and Si. In other words the reaction leading to the attainment of equilibrium at 1025° C is prevented as the ordered Al-Si arrangement cannot pass into one with measurable disorder, the required activation energy being very high.

The Al-Si order-disorder boundary line A B in LAVES (1952, p. 561) equilibrium diagram under very slowly cooling conditions is above the two phase boundary curve CDE. If this picture is correct, we should expect only one phase by heating for 96 hours a mixture, say of composition Ab 50% Mi 50% since according to the diagram (Fig. 1) Al-Si order-disorder transformation takes place at an appreciable rate at a temperature (T_1 say) higher than that (T_2 say) required for the homogenisation of the mixture, for the system will be arrested only at the higher temperature T_1, where there is no barrier for Na — K homogenisation. If the attainment of equilibrium with respect to Na — K distribution takes longer than 96 hours, the $d_{20\bar{1}}$ values for both the phases should change on longer heating; but the constancy of $d_{20\bar{1}}$ values found after heating for a period much longer than 96 hours indicate that equilibrium with respect to Na — K distribution is reached.

This induces me to propose the following equilibrium diagram for alkali feldspar. (See fig. 2). In this diagram, the Al-Si order-disorder boundary is not a straight line but is represented by the curve ADEB, the portion AD having a negative slope. BE having a positive slope and the curve truncates the top of the miscibility curve CDEF for the microcline — albite mixtures. It is probable that AD has a negative slope as shown in figure 2; since the microcline-disordered K-feldspar transition point may be lowered by the presence of a foreign phase (albite), even though the foreign phase enters into solid solution, as the entry of Na atoms in the potassium positions will cause a strain lowering the latter's strength and stability so that the inversion be-

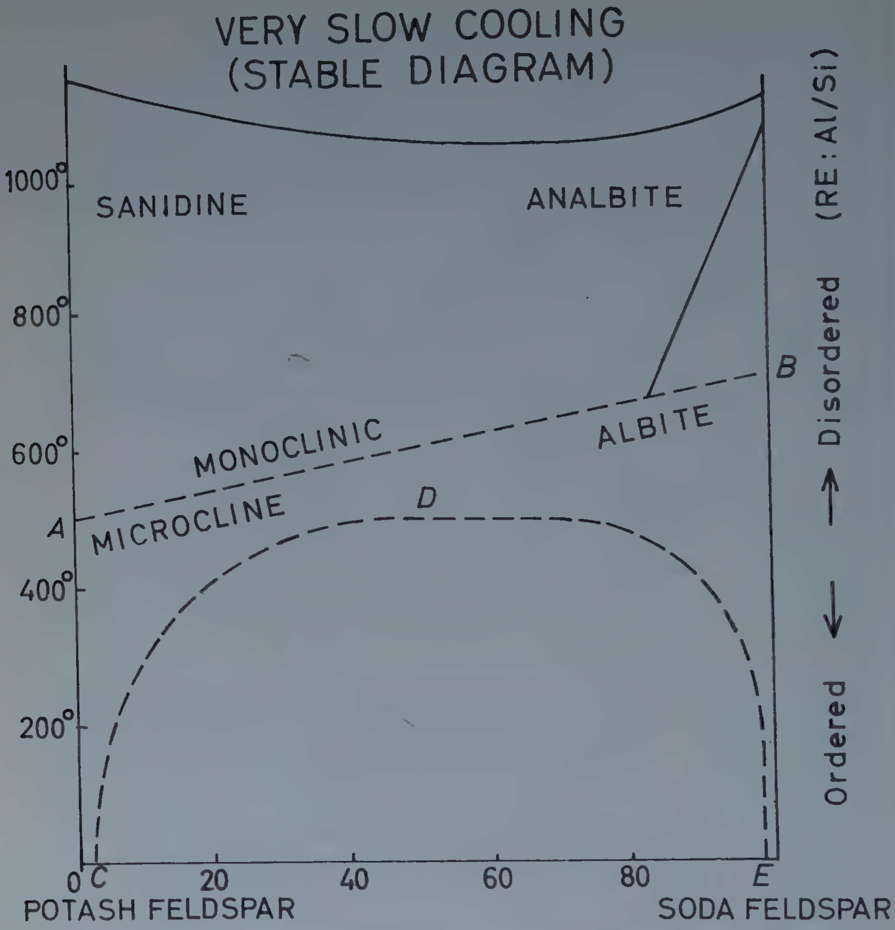

Fig. 1. Phase diagram of alkali feldspars. (After LAVES, 1952, p. 561).

comes possible at a lower temperature. The order-disorder line ADEB will therefore have the same form (passing through a minimum) as the solidus line in the system $KAlSi_3O_8$ — $NaAlSi_3O_8$ at such conditions where leucite field is absent. CDEF represent the two phase regions of the ordered alkali feldspars, which may extend metastably above the line DE as shown by the dotted line DD'E. The curve DHKE full line) shows the miscibility gap of the completely disordered feldspars (sanidine — analbite) worked out by BOWEN and TUTTLE (1950): this curve too may extend metastably below the line DE as

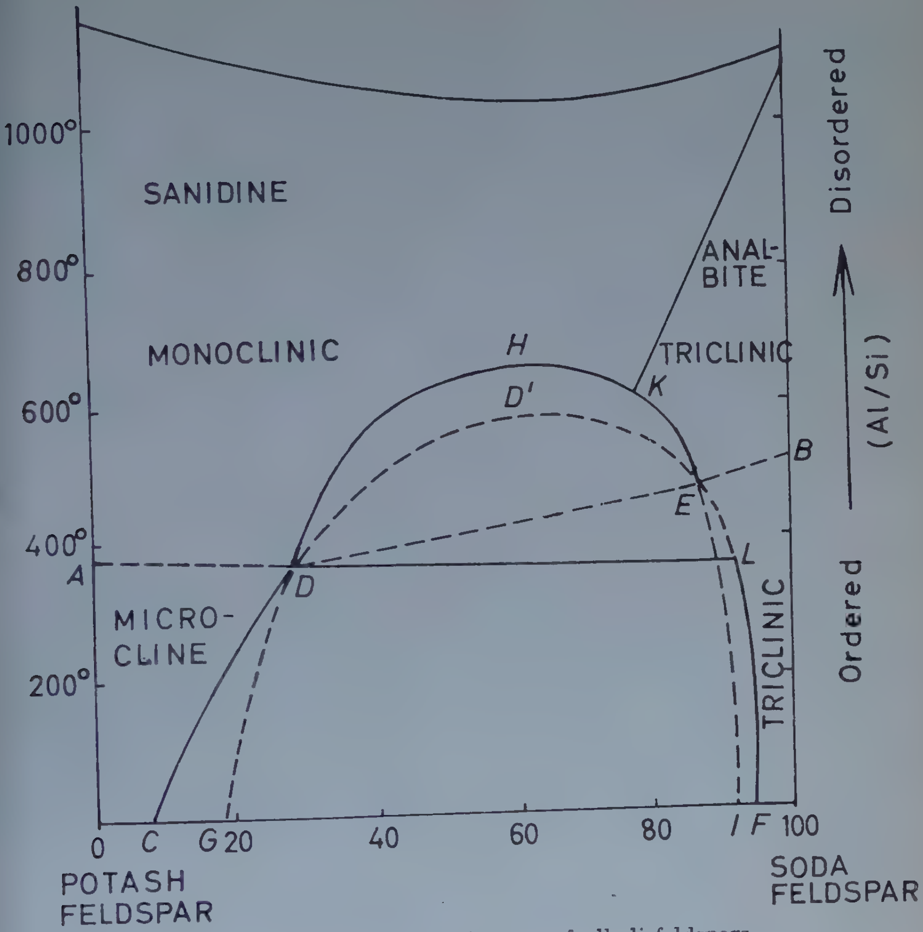

Fig. 2. Probable phase diagram of alkali feldspars.

shown by the dotted segments DG and EI. The JK represents the monoclinic/triclinic displacive transformation line which was proposed by Laves (1952). The Al-Si order-disorder line AB (points for it) has been taken arbitrarily at temperatures below 500° C and 700° C respectively as these should be lower than T_c values for the same as determined by Goldsmith and Laves (1954) and Bowen and Tuttle (1950) respectively. The diagram proposed is entirely qualitative and the temperatures shown may be fictitious. The main points that I want to indicate are: 1) that the Al-Si order-disorder line truncates the top of the miscibility gap for microcline-albite mixtures and 2) that the Al-Si order-disorder line is not a straight line but has probably the form shown by the curve ADEB.

It will be seen from table 1 that complete homogenisation is possible (within 96 hours of heating) in mixtures containing up to a little less that 30% albite. With mixtures with 30% albite to about 90% albite two phases appear. This leads me to postulate that the Al-Si order-disorder reaction barrier prevents further homogenisation when the mixture attains a temperature corresponding to DL or higher. In other words, the K-Na interchange between the two phases D and L is nearly arrested. The slight change in the position of $d_{20\bar{1}}$ line for the phases is due to progress of homogenisation metastably along the line DD'EL for the ordered potash phase, consequently affecting also the Na phase.

The reality of the line DD'E is that it represents a metastable state of equilibrium attained in cases of rapid heating.

Certain features of naturally occurring microcline-perthites and orthoclase perthites can be explained better with the help of the proposed diagram than with the Laves diagram. In nature, we rarely find microcline microperthites containing more than 40% albite. The Laves diagram (Fig. 1) (Laves 1952, p. 561) indicated that a triclinic microclinic feldspar containing more than 40% albite in solid solution can exist as a single phase if the temperature is above that indicated by the phase boundary line CDE; such a phase when cooled can easily separate into two phases giving rise to microcline microperthites when the temperature comes within the field CDE. If the Laves diagram is true, microcline perthites with more than 40% albite should not be so rare. But if the proposed diagram is true (Fig. 2), there is no way of getting microcline perthites with more than 30%—40% of albite, as in all such cases the alkali feldspar would exsolve prior to ordering resulting in various types of orthoclase (disordered) perthites. Since Al-Si ordering is very sluggish, in most cases, the exsolved phases cannot attain any further appreciable ordering after this phase separation. The temperature at this region is low enough to induce such ordering. However it is quite possible for an alkali feldspar of any composition to form metastably orthoclase perthites due to exsolution taking place earlier than (or with the absence of) ordering, thus overlapping the field of microcline perthites.

Effect of Soda Content on Triclinicity Values

It has often been stated that the triclinicity values determined from the different $d_{131}-d_{1\bar{3}1}$ is a measure of the degree of Al-Si disorder in the potash feldspars. Goldsmith and Laves, 8, 1954) have pointed out that the entry of sodium ions in a potash feldspar lattice would also change the triclinicity values. This has been observed by me in the case of mixture of microcline and albite heated at 1025° C ± 3 for 96 hours. The change in the triclinicity values measured from $d_{131}-d_{1\bar{3}1}$ for a number of these mixtures are given below to substantiate the above statement.

	Triclinity value
Microcline	.94
— with 10% Ab	.85
— with 20% Ab	.76
— with 30% Ab	.76
— with 40% Ab	.74

As it has been already indicated that practically no disorder of Al-Si sets in by the thermal treatment done on the above samples. The change in triclinicity values appears to be entirely due to replacement of potash ions by sodium in the crystal.

Two naturally occuring "orthoclase" microperhtites have been similarly heated at 1025° C ± 3 for different intervals of time and the changes in the triclinicity values as measured by the differences between 131 and $1\bar{3}1$ are recorded and given in the table 2 as follows:

Table 2. Changes in triclinicity values (as expressed by the width of 131 doublets doublets) on heating orthoclase perthites.

Sp. no.	Original width of 131 doublet	3h	6h	8h	24h
67 (Or: 65, Ab: 35[1])	.154 cm	.150	.143	.135	.135
70 (Or: 55, Ab: 45[1])	.171 cm	.170	.149	.143	.137

[1] The values are approximate, obtained from chemical analyses of Nordmarkite feldspars collected from Gjelleråsen area, Oslo, Norway. (The x-ray powder patterns are taken by a Nonius Guinier Camera of effective diameter of 22.98 cm).

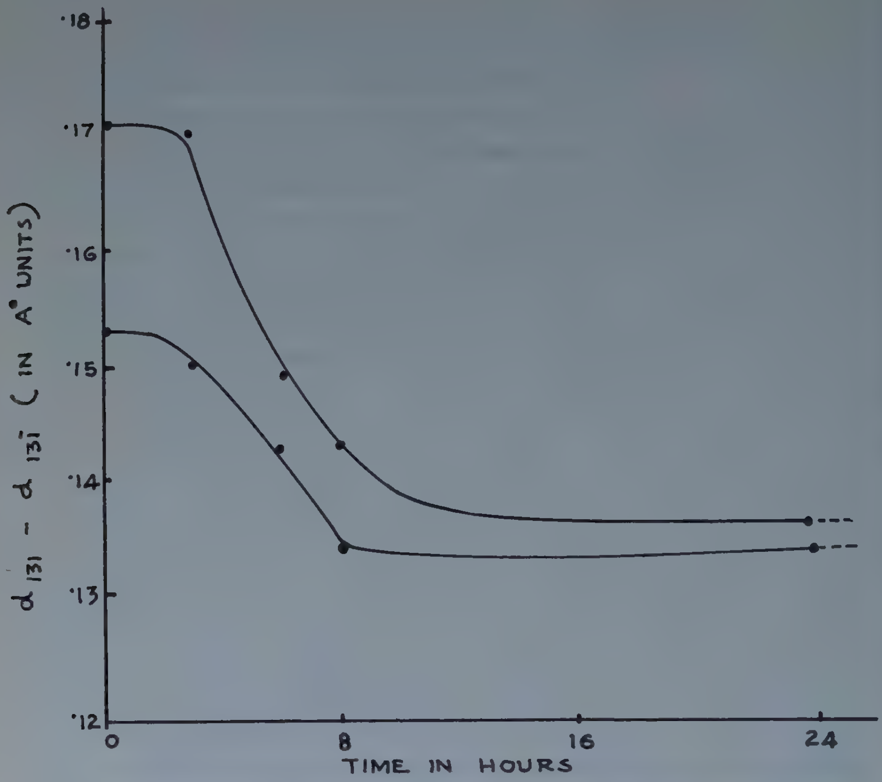

Fig. 3. Homogenisation and its effect on the triclinicity value or potash feldspars in some perthites.

These results are represented graphically in figure 3. These studies indicate the fact that homogenisation, as expressed by changes in doublet values, show a certain time lag for the initiation of the homogenisation. It will be seen from the figure that during the first 2 hours or so, the triclinicity values suffer no change, and during the next 6 hours there is a large change in the triclinicity values. After this period further heating does not effect the triclinicity values any more and the $d_{20\bar{1}}$ peak for othoclase — analbite phases coalesces into one clear line by this time. The first 2 hours may be regarded as a period of incubation for sodium and potassium ions to be released from their lattice position and to move through the crystal by jumping from one lattice vacancy to another in the same crystal in which they are originally present. After the period of 8 hours, since the homogenisation is complete there is no further change in the triclinicity values. Rate of homogenisation in all perthites is not alike though the process appears to be the same. The grain size of the perthitic components is one of the factors that has got to be taken into consideration.

Classification of Perthites

Many attempts have been made to associate a genetic significance with the occurence of the perthites in various rocks. The most recent studies have been made by TUTTLE (1952) and MACKENZIE and SMITH (1955, 1956). The former relied on optic axial angle and divided the alkali feldspars into four series, the latter two authors have attempted to correlate X-ray structural features with the divisions proposed on the basis of optic axial angles. Mackenzie and Smith have also proposed a simple method to distinguish low and high soda phases under single crystal X-ray studies. On the basis of their X-ray study they proposed a division into two series: sanidine cryptopethites which under X-ray show a high soda phase, and orthoclase microperthites which show a low soda phase.

To give a general division of the perthites, I suggest the following:

Perthites with disordered Al-Si distribution. (*High perthites*). In these either one or both of the components are partially or fully disordered. Those with a disordered potash phase and disordered soda phase may be classified under this group. This is equivalent to the sanidine-cryptoperthites of Tuttle and Mackenzie. Either as a part of it or as another group, we may recognise perthites, in which the potash feldspar component is disordered and the soda feldspar is more or less ordered. These may be called *intermediate perthites* which corresponds to the "Orthoclase-microperthites" of Tuttle and Mackenzie.

Perthites with ordered components. (*Low perthites*). In these both the soda and the potash feldspar components are fully ordered and this series is equivalent to "microcline-microperthite" series.

Acknowledgements

This work has been carried out at the *Geologisk Museum,* Oslo under the guidance of Professor Tom. F. W. Barth. I am thankful to Mr. Olav Christie for his help in heating the samples. I thank Professor Barth, Oslo, and Professor S. K. Siddhanta of this Institute for their help in the presentation of this paper.

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STRUCTURAL GEOLOGY OF THE SOKUMVATN AREA, NORTH NORWAY

BY

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Abstract. The Sokumvatn Area is representative of the 'deeper' part of the Caledonian orogenic belt in N. Norway. The general structure is a synform with a N—S axis (the Sokumvatn Synform); the Harefjell Gneissic Group, dioritic to quartz-monzonitic in composition is surrounded on three sides by the Sokumfjell Marble Group, composed predominantly of saccharoidal marbles. Both groups are heavily veined by granites.

An important slide zone occurs between the two main rock groups and other slides occur within each group. The sliding is related to large scale overfolds on E.N.E.—W.S.W. axes which may be recognised in both groups. These overfolds have been refolded in the Sokumvatn Synform so that they now have nearly vertical plunges in its steep eastern limb.

Minor structures associated with the two fold episodes differ in style. Those in the gentle western limb of the Sokumvatn Synform belong mainly to the more plastic first folding while those in the steep eastern limb are largely due to the less plastic second folding. The attitude of the second fold structures is variable and dependent on the attitude of the structures on which they were superimposed.

The evidence from the metamorphic history and granite emplacement is briefly discussed to emphasise the difference in age between the two fold episodes. The first folding was synchronous with the regional metamorphism while the second folding occurred during a period of granite emplacement with falling temperature.

Introduction

The Sokumvatn Area lies inside the Caledonian orogenic belt in Norway well to the west of the zone of post-metamorphic thrusting which has been recognised in the Sulitelma area close to the Swedish frontier (e.g. KAUTSKY 1953, see map, fig. 1). It is part of a much larger area which is under study by a research group from University College London. The geology of this larger area and its integration into the Caledonides, will be presented at a later date: the present paper is concerned mainly with the evidence for two periods of folding, on different axes, at distinctly different times. Although cross folding has been widely recognised in Norway (e.g. LANDMARK 1951), it has not previously been analysed from a basis of large scale mapping.

The only previous work in the Sokumvatn Area has been on a reconnaissance scale by the Norwegian Geological Survey (N.G.U.) and especially by REKSTAD (1912, 1913). HOLMSEN (1932) has summarised the available information in the description accompanying the N.G.U. 1:250.000 map of Rana. The present mapping has been on the scale of aerial photographs (approx. 1:16.000) and locally on larger scales.

The mapping was carried out between 1953 and 1955 and was shared with K. J. Ackermann, B.Sc., who has kindly placed his results at the writer's disposal for this paper. Ackermann has been chiefly concerned with petrological study of the meta-sediments of the area while the writer has dealt with the structural geology and with the petrology of a meta-volcanic group (Ph. D. Thesis, 1958). The writer is personally familiar with all the area except a portion of the south-eastern quadrant, where he has relied entirely on Ackermann's observations. A preliminary account of the results contained in this paper was presented in a lecture to the Geological Society of London in May 1956.

The area is particularly suitable for the demonstration of double folding for two main reasons. Firstly, there are present several con-

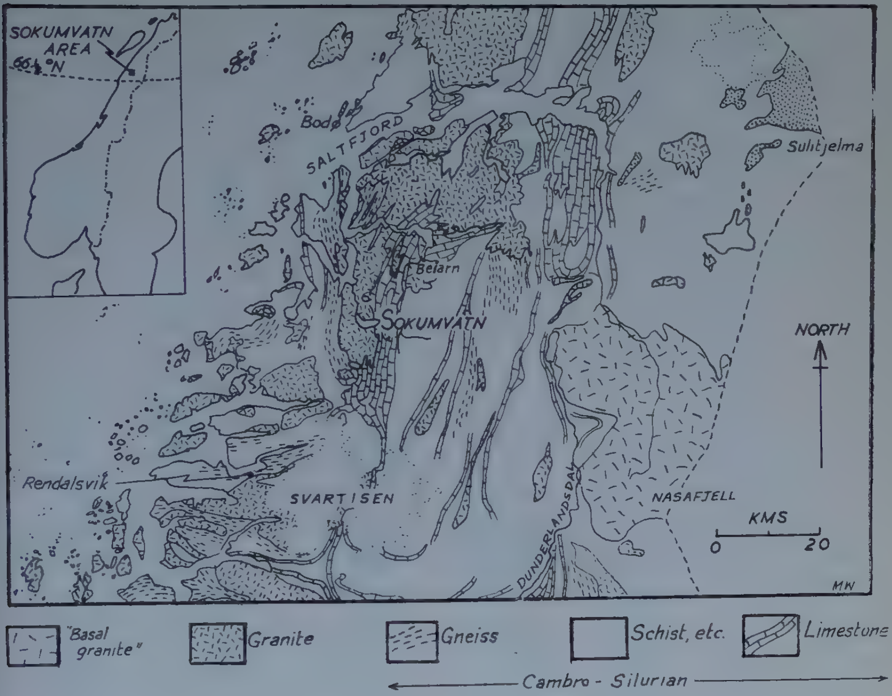

Fig. 1. Geological setting of the Sokumvatn Area.

trasting lithological types which allow ready determination of the geometry of the structures; and, secondly, the area is a locus of granite veining and granitisation, a phase of which facilitates the separation of the fold episodes in time.

Two main rock groups are present and both are excellently exposed. The Harefjell Gneissic Group, dioritic to quartz-monzonitic in composition, is surrounded on three sides by the Sokumfjell Marble Group (Plate 1). The petrology of the Gneissic Group will be discussed elsewhere, but it must be mentioned here that it is interpreted as a series of pre-orogenic volcanic rocks, which has been folded, metamorphosed and metasomatised to produce the present assemblage. Most critical to this interpretation is the structural evidence which demonstrates that this is a layered series, interbanded with subordinate meta-sediments, mainly marbles with impure quartzites. These marble bands allow the recognition, in the Gneissic Group, of major overfolds

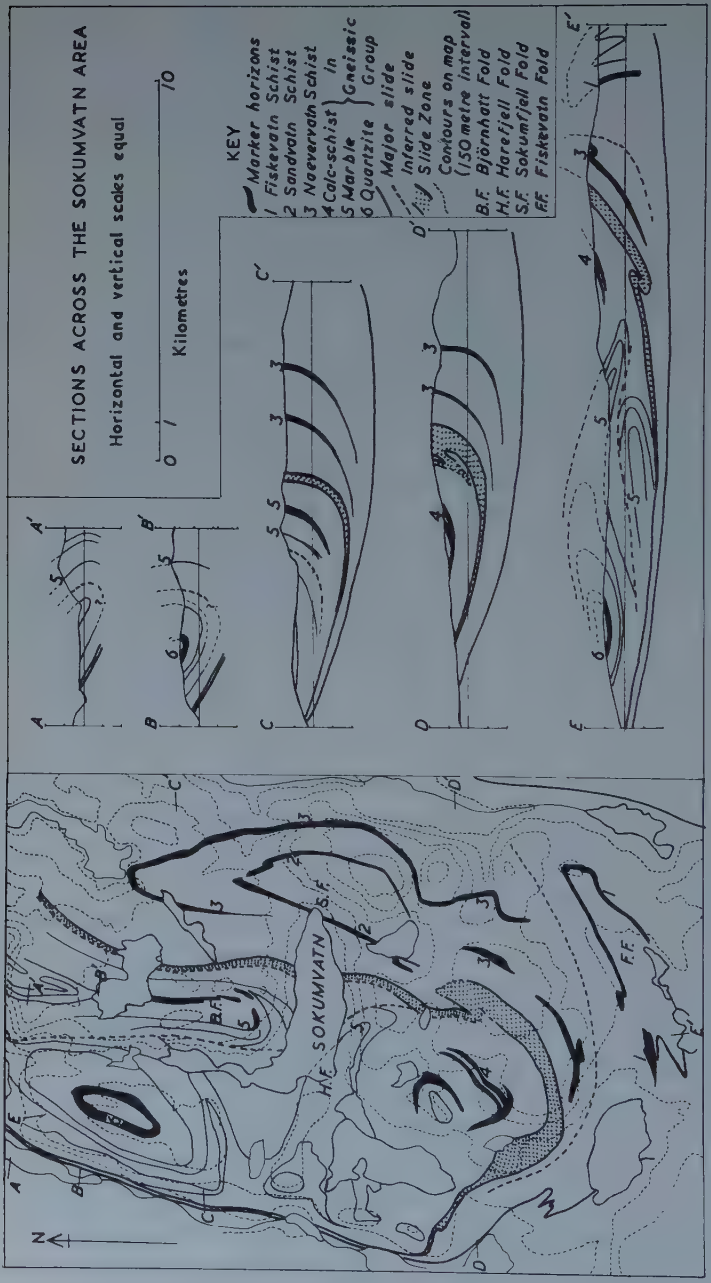

Fig. 2. Geological sections across the Sokumvatn Area.

analogous to those which are even more clearly shown in the Marble Group by bands of pelitic schist with subordinate calcsilicate rocks. It can be shown by reference to these fold structures that there is a strong 'structural unconformity' between the Gneissic and Marble groups, and a zone of sliding, the Staburfjell Slide Zone, is recognised between them (fig. 2).

Mapping of the contrasting bands, in both groups, also allows the determination of the 'sheet dip'[1]. Thus it is often possible to determine the attitude, and especially the plunge, of the major structures, independently of the minor structures. Minor structures can then be interpreted in relation to the known geometry of the major folds and are not used to infer it.

The general nature of the structural problem is indicated by the plots of linear structures (B diagram) and of poles to bedding and banding (πs diagram) in equal area projection (figs. 3a & b). The area as a whole is emphatically not homo-axial. The available data have been examined by modern methods in smaller units (fig. 3) but they are inadequate for full graphical statistical analysis.[2] Smaller units are less heterogeneous than the area as a whole, but only in the western part of the area is there a close approach to homo-axiality (fig. 3f).

Such structural complexity could be due to a number of factors. Basement interference (e.g. KVALE 1948) or igneous intrusion (e.g. LYONS 1955) could cause departures from a homo-axial pattern; but in the Sokumvatn Area the relations can be interpreted in terms of two periods of folding (cf. WHITE and JAHNS 1950, RAMSAY 1956).

[1] It is necessary to distinguish between the sheet dip, which is the dip of a lithological unit as a whole, and the average dip which can be determined from statistical analysis of individual measurements taken in the field. Where minor folding is present these will not be the same. The sheet dip is best determined from the relationship between trend of the outcrops and variations in topography. Individual measurements of the bedding within the lithological units are also necessary to decide whether changes of trend are, in fact, due to topographic effects or to real changes in strike.

[2] Such data is much less easy to acquire in a region of basic igneous gneisses and marbles such as this than in pelitic schist terrains, and, in any case, time was limited. But in the writer's opinion the direct method of determining the attitude of major structures which was possible in this ground, is preferable to graphical statistical analysis, since the structures analysed may not be related to the major structures on which they are superimposed.

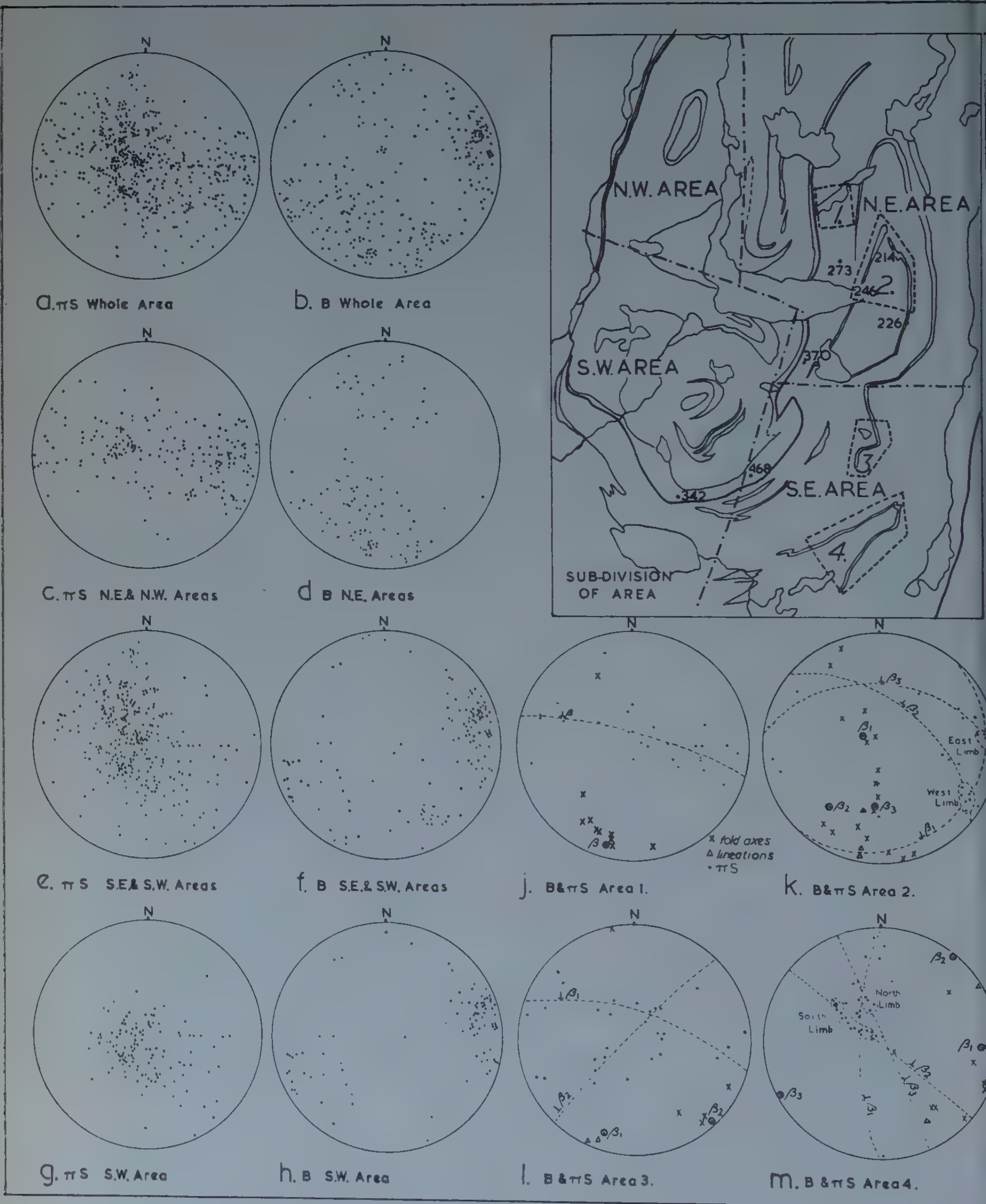

Fig. 3. Equal Area projections of structural data.

During the earlier period of folding, overfolds have been produced on axes trending W.S.W.—E.N.E., in both the Gneissic and Marble Groups. During the later period of folding, these have been refolded on nearly N—S axes.

Both the Marble and the Gneissic groups have been heavily veined by granites, and there has been important granitisation in the Gneissic group. In some places these phenomena have obscured the tectonic relations but they have also served to separate the two fold episodes in time. The first folding took place during the main Regional Metamorphism (amphibolite facies), while the second folding took place during the subsequent granitisation under very different conditions, when both the temperature and hydrostatic pressure were probably much lower.

One major difficulty in the structural interpretation of the area is the lack of any evidence of the original stratigraphic succession. General comparison with other parts of Norway suggests that the rocks are probably Ordovician, but there are no fossils, nor any sedimentary structures, such as current or graded bedding, to indicate the original order of superposition. This does not affect the determination of the geometry of the major structures, or the recognition of double folding, but it does make it difficult to assess the importance of the main slides.

The Staburfjell Slide Zone

The contacts of the rocks of the Sokumvatn Area with the schist groups to east and west are important slides, the discussion of which is beyond the scope of this paper. However there has also been a great deal of syn-metamorphic sliding within the area, the most important of which has taken place in the Staburfjell Slide Zone between the Gneissic and Marble groups (see map, fig. 2 and Plate 1).

The Staburfjell Slide Zone forms a distinctive lithological unit generally with nearly parallel sides. The rocks within the slide zone consist of a mixture of dioritic gneisses and meta-sedimentary schists in variable, but often roughly equal, proportions. The gneisses are like those of the Gneissic Group, and are regarded as meta-volcanic rocks, while the schists are mainly pelitic with subordinate calcsilicate types and rare marble bands. Thus there is a marked contrast with both the

Marble and the Gneissic groups. The Marble Group consists of dominant marbles with subordinate meta-sedimentary bands: the Gneissic Group consists of dominant meta-volcanic gneisses; narrow marble and impure quartzite bands are present but, except in the southern part of the area north of Staburfjell, pelitic and calc-silicate shists are absent. It is tempting to suggest that the Slide Zone is a distinct stratigraphic unit and that it may represent the original passage beds between the Gneissic and Marble groups.

This lithological contrast makes the contact of the Slide Zone with the Marble Group the most prominent line on the map in the whole area. The contact with the main part of the Gneissic Group is also easily recognised where it brings meta-sediments in the Slide Zone against the rocks of the Gneissic Group or where it is marked by a marble band in the Gneissic Group. Locally however, this contact is mappable with less precision when dioritic gneisses occur on both sides, as, for example, immediately north and south of Sokumvatn, where there is also considerable post-tectonic granitisation to obscure the tectonic relations.

The Slide Zone varies from a maximum thickness of the order of 400 metres to places where it has been completely eliminated by the sliding between the Marble and Gneissic groups. Between Little Sokumvatn and Urfjell on the eastern side of the Gneissic Group the thickness is about 200 metres. On Urfjell this thickness may be almost doubled where the Slide Zone makes a projection into the Marble Group (the Urfjell 'bulge'). Between Urfjell and Staburfjell the zone is about 300 metres thick but it thins north-westwards so that it is only about 75 metres thick in Langvassa, and it disappears altogether as a mappable belt between Storvatn and Samelvatn so that the Marble and Gneissic groups are in direct juxtaposition on a planar slide contact. East of Samelvatn it reappears and runs northwards with a thickness of about 20 metres.

The contacts of the Slide Zone both with the Gneissic Group and the Marble Group must be regarded as major slides since they cut across the major structures in the two groups. The Gneissic Group can be divided into three major units, the Harefjell Fold, the Bjørnhatten Fold, and the area north of Staburfjell; each of these units comes into contact with the Slide Zone. In the Marble Group, the Nævervatn Schist which is over one and a half kilometres from the

Slide Zone at Navnløsvatn approaches to within one hundred and fifty metres of it south of Staburfjell, despite the fact that the dip is much gentler at the latter locality. Thus there is a thinning of about ninety five per cent in the western limb of the Sokumfjell Fold, yet the thickness of marble between the Nævervatn Schist and the nearby dolomitic marble remains virtually the same south-east of Staburfjell as at Navnløsvatn. It appears that the western limb of the Sokumfjell Fold is not in reality thinned by attenuation to any marked extent, but is progressively eliminated against the Staburfjell Slide Zone. This is particularly obvious on the map (Plate 1) in the area W.S.W. of Sandvatn, where a thick quartzite band strikes into the Slide Zone and must be cut out against it, though the actual point of contact was not observed.

Within the Slide Zone no constant succession can be recognised. On Staburfjell the zone is divisible into three main units, but these cannot be recognised elsewhere. Individual bands can be mapped along the zone for only short distances, and no two traverses across the zone give the same succession. Either meta-sedimentary or meta-volcanic rocks may lie on the contacts with both the Gneissic and the Marble groups and there is clear evidence that these relations are tectonic.

Minor overfolds of similar style and attitude to those associated with the major overfolds on E.N.E.—W.S.W. axes in the Marble and Gneissic Groups are frequent (fig. 7c). During the folding and sliding the more massive meta-volcanic gneisses have been disrupted to 'float' in the less competent pelitic schist. On a smaller scale a similar phenomenon is shown by the calc-silicate bands in the pelitic schists. Numerous localities show fragments of calc-silicate rock scattered in the homogeneous schists (fig. 7d and e), in a manner resembling xenoliths in an igneous intrusion. The fragmentation is often irregular; but it does take place in directions parallel to the fold axes and lineations. Such extreme deformation is confined to the Staburfjell Slide Zone. It is not present in either of the main rock groups although both have suffered intense deformation with the production of isoclinal overfolds.

Thus there is no doubt that the Staburfjell Slide Zone is a zone of great differential movement. The concordance of the minor structures in the zone with those in the overfolds outside it shows that the sliding took place in the same period of deformation as the overfolding. There are no cataclastic structures to indicate post-metamorphic deformation and the fabrics show that crystallisation (in the Amphibolite Facies) outlasted deformation. On the other hand the relatively brittle behaviour of the calc-silicate bands suggests that they were already metamorphosed at the time of the maximum deformation. It follows that the sliding and overfolding were synmetamorphic.

The actual amount of relative displacement between the Gneissic Group and the Marble Group cannot be assessed because their stratigraphic relationship is unknown. It has been suggested above however that the Slide Zone may represent the original passage beds between the two groups and therefore that they were originally in juxtaposition. If so, the Slide Zone is clearly to be regarded primarily as a zone of intense differential movement which has allowed disharmonic folding to take place between the rock groups above and below it.

The Sokumvatn Synform

The Staburfjell Slide Zone provides an excellent horizon for the recognition of the major structure of the area, marking as it does the boundary between the two major rock groups. The sheet dip can often be determined and it indicates that the zone is disposed in a large synformal structure — the Sokumvatn Synform — with a gentle western limb and a steep eastern limb.

In the extreme north-east of the area, north of Lille Sokumvatn, the Slide Zone is nearly vertical, but everywhere else it dips under the Gneissic Group. On the eastern side of the Gneissic Group, from Lille Sokumvatn to Staburfjell, the dip is steep in a westerly direction. This is best seen on the map from the change in trend of the outcrop caused by the deep topographic depression in which Sokumvatn lies. On the western side of the Gneissic Group, from Sundsfjord to Staburfjell, the dip is gentle to the east. The outcrop of the Slide Zone swings around Staburfjell as the dip changes from gentle to the north-east, at the western end, to steep to the northwest at the eastern end. South of Staburfjell the gentle northly dip indicates that the synform has a northerly plunge.

One feature, the Urfjell 'bulge', slightly complicates this picture of a simple synform in the Staburfjell Slide Zone. As a result of the northernly plunge of the synform the outcrop of the Gneissic Group should widen northwards from Staburfjell. It does so as far as Urfjell but there it suddenly narrows beyond the Urfjell 'bulge'. It may be inferred therefore, that the effect of the general northernly plunge has been suddenly counterbalanced and that a lower structural level in the synform has been brought up immediately north of Urfjell. Examination of the map suggests that this is due to a cross structure which affects not only the Staburfjell Slide Zone but also the rocks of the Marble Group east of Urfjell: the limbs of the Sokumfjell Fold are deflected round the Urfjell 'bulge' (fig. 11). A suggestion that the structure also carries over to the western side of the Gneissic Group is to be found in a much smaller bulge in the Staburfjell Slide Zone east of Storvatn. It has not been possible to work out the details of the structure owing to lack of exposure and abundant granite veining on Urfjell but its general nature as an E—W antiformal structure, which has the effect of cancelling the northerly plunge of the Sokumvatn Synform, is illustrated in fig. 2 (section EE^1).

It may be conjectured that the structure was formed simultaneously with the main synform. The northerly plunge of the latter may be an inherited effect from a general northerly dip of the rocks before folding. E—W cross folds may then be regarded as a necessary accompaniment of the main N—S fold in order to allow the maximum strain to take place, not normal to the northerly dip, but in a vertical plane, which is likely to have contained the maximum and minimum principal stresses.

In spite of the complication caused by the Urfjell 'bulge' it is evident that the Staburfjell Slide Zone has been folded into a broad asymmetric synform with a N—S axis. But as explained above, the intense deformation within the Slide Zone has been produced at the same time as major isoclinal overfolds in both the Marble and Gneissic groups. These overfolds are described later when it is shown that they were produced on E.N.E.—W.S.W. trending axes earlier than the Sokumvatn Synform. At this point it is only noted that their presence complicates the outcrop pattern within the main synform; but if only the sheet dips are examined it becomes evident that both the

Marble and Gneissic groups share in the general synformal structure which is so easily recognised in the Staburfjell Slide Zone.

The axis of the Synform runs northwards from Staburfjell, through the eastern end of Sokumvatn and roughly along the line of the stream which drains Sokumvatn to Holmsundsfjord. West of this axis are gentle dips, usually easterly. There is a slight east—west culmination, which gives a northerly component to the dips on its northern side and a southerly component on its southern side. Consequently the rocks highest in the structural successions occur in the neighbourhood of summits 635 (east of the south end of Langvatn) and 495 (west of Lille Sokumvatn) (fig. 11 and Plate 1).

East of the axial line defined above, are steeper dips to the west. South of Sokumvatn the dips are gentle to moderate near the axis, and only exceed 50° or 60° on the extreme eastern margin of the Gneissic Group. Immediately north of Sokumvatn the whole of this eastern limb of the synform has steep westerly dips and at Lille Sokumvatn it is about vertical. Further north still it becomes overturned so that on Bjørnhatten it has a moderate dip to the east.

The Sokumvatn Synform thus has an approximately north—south axis, and is an open fold in the south and an overturned fold in the north. The form of the fold is indicated in the sections (AA′—DD′) of figure 2. The πs diagram for the northern half of the area (fig. 3c) confirms this general simple structure.

In the west of the area, the narrow outcrop of the Marble Group dips gently to the east under the Staburfjell Slide Zone, and may be regarded as part of the western limb of the Sokumvatn Synform.

On the east of the Gneissic Group, along the eastern end of Sokumvatn, the sheet dip in the Marble Group is steep to the east and this part of the Marble Group may be regarded as part of the eastern limb of the Sokumvatn Synform. However, further east, the sheet dip becomes vertical, in crossing the core of the Sokumfjell Fold, and then easterly in the eastern limb of that fold. Thus superficially, the Sokumfjell Fold has the appearance of a complementary antiform to the Sokumvatn Synform. This, in fact, is not correct as will be shown when the Sokumfjell Fold is described. The whole of the Marble Group between the Gneissic Group and the Vegdal Schist Group is properly to be regarded as part of the eastern limb of the Sokumvatn Synform (section DD′ fig. 2).

No continuation of the Sokumvatn Synform can be recognised southwards from Staburfjell. Between Fellvatn and Arstadvatn the strike varies from easterly to north-easterly as is clearly shown by the outcrops of the Fiskevatn Schist, and there is no indication of any major folding on north—south axes. The πs diagram for the southern half of the area (fig. 3e) shows a different and more complex pattern than that for the northern half (fig. 3c), with a suggestion of a main girdle with an E.N.E.—W.S.W. axis.

Folds in the Harefjell Gneissic Group

Four main rock types have been mapped in the Gneissic Group, viz. massive diorite, foliated dioritic rocks, quartz-monzonite gneiss and granite. The compositional variations are largely due to post-tectonic granitisation, and it is probable that all the rocks were originally broadly 'dioritic' in composition. But the distinction between massive diorites and foliated diorites probably represents an original lithological difference, and the meta-sedimentary bands, which are so valuable in determining the structure of the group, are never found in the massive diorites.

Detailed mapping of the meta-sedimentary bands in the foliated diorites shows that although they are often broken, they follow definite horizons which are generally parallel to the banding and schistosity, and to the major contacts, in the Gneissic Group. The bands are generally parallel sided and less than ten metres thick, though one or two are over fifty metres thick.

Within the bands, however, although the bedding is generally parallel to the contact, there is evidence of extreme deformation, and isoclinal folds are common (fig. 7a and b). Thus the planar contacts of the bands, parallel to the schistosity in the adjacent gneisses, are obviously tectonic. Eash band is clearly a slide zone between the more massive rocks of the Gneissic Group on either side.

In spite of this deformation, individual bands may often be followed without a break for several hundred metres, and one band south of Lille Sokumvatn can be traced for fully four kilometres before it disappears beneath that lake. It may be inferred that before the deformation the bands were fully conformable members of the layered sequence, and much more continuous than now. As mapped, the bands define the major structures within the Gneissic Group, and two major folds are revealed, here called the Bjørnhatten and Harefjell folds, north and south of Sokumvatn respectively (fig. 2).

Unfortunately the meta-sedimentary bands are not individually distinctive, so that correlation is generally only possible by continuity of outcrop or feature. Therefore they cannot be used to correlate the successions in the two folds. However, the massive diorite is distinctive and occupies about the same position in both folds. Its outcrop is in fact continuous from the one fold to the other. Thus the evidence favours a general correlation of the rocks of the two folds.

The gneisses between Lake 545 and Staburfjell cannot be correlated with those further north. The meta-sedimentary bands in these gneisses consist of dominant pelitic schists with subordinate calc-silicate rocks, while those of the Bjørnhatten and Harefjell Folds consist almost exclusively of marble and impure quartzite.

THE BJØRNHATTEN FOLD

The Bjørnhatten Fold lies in the steep eastern limb of the Sokumvatn Synform, north of Sokumvatn. The trike in both limbs is about north—south. In the extreme north, on Bjørnhatten, both limbs have a moderate dip to the east. Southwards the dip steepens, so that on either side of Lille Sokumvatn the limbs are nearly vertical, and towards Sokumvatn, both limbs dip steeply to the east. Thus the fold is everywhere isoclinal, and the attitude of the limbs depends on the position in the main Sokumvatn Synform. When the latter fold is overturned, the limbs of the Bjørnhatten Fold dip east: where it is open, they dip west.

The plunge of the Bjørnhatten Fold can be determined approximately from the sheet dip round the fold closure. On the north shore of Sokumvatn the sheet dip is fairly steep right round the fold closure, and it is estimated that the plunge must be about 50° to the south-west. On the north shore of Lille Sokumvatn the sheet dip is less clear, but the plunge appears to be extremely steep.

The intense deformation which occurs in the meta-sedimentary bands has taken place on the same axes as those indicated above for the major structure. Thus the lineations and axes of overfolds in these bands also plunge at about 50° to the south-west, north of

Sokumvatn, and are nearly vertical at Lille Sokumvatn. It seems clear therefore that these minor structures are the result of the formation of the Bjørnhatten Fold, with which they agree in style.

In addition there are also other minor folds present which have almost horizontal axes trending north-south, parallel to the axis of the main Sokumvatn Synform. These folds are generally cylindroidal with amplitude and wavelength about equal, in marked contrast to the isoclinal overfolds with steep plunges.

THE HAREFJELL FOLD

The Harefjell Fold is a much less obvious structure than the Bjørnhatten Fold and its status is less certain. The core of the fold can be distinguished on the steep slope rising from Sokumvatn to the west end of Harefjell. There, a marble band can be followed right round the fold closure. In both limbs this band dips gently to the south-south-west. Intense overfolds within the band (fig. 7a) plunge gently to the west-south-west and this appears to be the plunge of the major overfold also.

Several other marble bands may be found in both limbs of the fold and these also show tight overfolds plunging W.S.W., but none is continuous round the fold closure. However the massive diorite can be traced round the fold. West of Sokumvatn it dips gently to the S.S.W. in the lower limb of the fold. Across the northern end of Langvatn and towards Harefjell it has about the same dip in the upper limb of the fold. The two limbs are linked on Harefjell where the dip steepens on the slope up from Sokumvatn.

Thus the Harefjell Fold is another isoclinal overfold and, like the Bjørnhatten fold, it plunges to the west-south-west.

The rocks west of Langvatn and those north of Lake 545 are probably to be regarded as part of the upper limb of the Harefjell Fold, but those south of Lake 545, with pelitic schist bands, are probably not. The latter also show fold axes and lineations with W.S.W. trend but the sense of movement, indicated by the folds, does not agree with that of the Harefjell Fold. It is probable that the fold is separated by a slide from this group of gneisses with pelitic schist bands. A slide has to be postulated in any case to account for the absence of this group elsewhere above the Staburfjell Slide Zone.

The rocks in the north-western part of the Gneissic Group appear to belong in the lower limb and core of the Harefjell Fold. At the south-west corner of Sokumvatn the quartz-monzonite gneiss dips gently S.S.W. over the massive diorite in the lower limb of the Harefjell Fold, and is overlain by the core of that fold. This same belt of gneiss dips gently to the N.N.W. above the massive diorite on the north side of the Stivatn culmination. Therefore it can be inferred that the rocks which overlie it to the north also form part of the core of the Harefjell Fold. It also follows that the Harefjell Fold overlies the Bjørnhatten Fold. This interpretation is illustrated in the section EE′ of figure 2.

It will be clear from the above description that the Bjørnhatten and Harefjell Folds and most of the minor folds associated with them have axes which make a large angle with the axis of the Sokumvatn Synform, and plunges which depend on the dips of the limbs of that Synform. Moreover they are isoclinal folds and the style of deformation which they exhibit is not readily correlated with the generally open structure of the Synform. It is therefore very difficult, if not impossible, to envisage that they were formed simultaneously with it. On the other hand, the hypothesis that they are early overfolds which have been refolded about the axis of the Sokumvatn Synform readily explains the observed facts. Yet stronger evidence in support of this hypothesis is to be found in the Marble Group.

Folds in the Sokumfjell Marble Group

The Marble Group forms an outcrop about 5 kilometres wide between the Gneissic Group and the Vegdal Schist Group to the east, and to the south it extends as a broad outcrop some ten kilometres beyond the present area (fig. 1). West of the Gneissic Group, however, the Marble Group is greatly attenuated. In the Langvassaa section it is about 400 metres thick, and between Langvassaa and Storvatn it rapidly thins to about 30 metres. North of Samelvatn it is still further reduced to less than 20 metres, but the marbles remain pure and rather massive. Towards the northern margin of the area these marbles pass into a similar thickness of impure marbles (bearing grossularite and vesuvianite), with bands of calc-silicate rocks, pelitic schists, and impure quartzites; amphibole bearing granite veins are also abundant. Deformation has been extreme and the marble shows many minor isoclinal folds, often with sheared out limbs.

Pelitic schist bands in the Marble Group demonstrate the existence of two major overfolds, the Sokumfjell Fold in the east of the area and the Fiskevatn Fold in the south. Similar bands are absent from the western area, but overfolding is evident from the form of the contact between the Marble Group and the Meløy Schist Group and from the smaller scale structures in both groups.

THE WESTERN AREA

Between Fellvatn and Storvatn, the Meløy Schists make numerous ridges trending W.N.W. These ridges have steep western slopes and gentle eastern slopes, corresponding to scarp and dip slopes respectively for the schistosity and bedding, which are nearly parallel and dip at about 30° to E.N.E. The scarp slopes sometimes reveal that the schistosity is, in fact, an axial plane schistosity to reclined isoclinal folds. The dip slopes show a prominent linear structure plunging down dip at about 20 to 30° between E. and E.N.E., and corresponding to the fold axes, to the intersection of schistosity and bedding, and to the dimensional orientation of mica. The scarp slopes mark prominent joints normal to the fold axes. The microfabric of these schists has not been studied, but in similar rocks to the south of Fellvatn, B. J. Walton (in a personal communication) finds that the quartz and mica fabrics are homotactic and show an 'ac' girdle normal to the fold axes. An analogous fold style and microfabric is also found in the schists of the Staburfjell Slide Zone in the western limb of the Sokumvatn Synform (fig. 9c—f).

South of Fellvatn more open folds were observed particularly in the more competent calc-silicate bands within the pelites. These folds are overturned to the south; the long northern limbs have a gentle north-easterly dip while the short, inverted southern limbs have a steep northernly dip. The fold axes and lineations plunge between 5° and 30° in a roughly easterly direction.

The attitudes of the structures in the Sokumfjell Marbles are quite comparable with those in the Meløy Schists. South of Staburfjell the general strike is W.N.W., but east of Staburfjell it turns

N.N.W., parallel to that in the Meløy Schists. The sheet dip is also similar to that in the schists and again it can be seen that isoclinal folds occur, plunging to the E.N.E. (fig. 7f).

The contact between the Meløy Schist Group and the Marble Group indicates the same style of folding on a larger scale. N.W. of Fellvatn, the large scale interdigitation of the two groups, with a constant north-easterly dip throughout, indicates that the contact has suffered isoclinal folding like that in the adjacent schists and marbles. South of Fellvatn a more open folding of the contact on similar axes is suggested.

It is clear therefore, that the rocks in this western area have suffered intensive overfolding on E.N.E. axes. This trend is similar to that of the overfolding in the meta-sedimentary bands in the Gneissic Group, which is of similar style, and which is associated with the major overfolds there.

In addition to these structures, however, there are others of different style and trend. In the Meløy Schists, the lineation and axial-plane foliation of the overfolds described above are sometimes seen to be refolded in very open structures whose axes have gentle plunges trending roughly N.N.W. The trend of the earlier lineation may change somewhat across these later folds. For example, in one case where the later fold was in the nature of a monoclinal warp of the earlier foliation, the earlier lineation changed its plunge from 45° at 76° E. of N. to 24° at 89° E. of N.

These warp structures are comparable both in axial trend and style with the main Sokumvatn Synform and they clearly suggest that the latter plays an analogous role on a larger scale. That is, the Sokumvatn Synform has refolded the earlier overfolds, including the Bjornhatten and Harefjell Folds, on a N—S axis, and thus it accounts for the steep plunges which they now show in its steep eastern limb. The available observations indicate a slight change in trend of the earlier structures in crossing the axis of the Synform. In the gentle western limb the linear structures generally trend between E. and E.N.E. but in the eastern limb there is a tendency towards S.W. rather than W.S.W. or W. plunges. This is readily explained by folding about a N—S axis.

The boudin axes of granitic minor intrusions also seem to be related to the later folding. In the Langvassaa section, fold axes and lineations have a very consistent plunge to the E.N.E., though a few warp structures occur on northerly axes. Planar granite intrusions are emplaced roughly parallel to the bedding[1], but their boudin axes have various trends between N.E. and N.W. and it is to be conjectured that these axes were developed by locally variable responses during the later movements.

THE SOKUMFJELL FOLD

General Geometry

In the western area described above the minor folds in the marbles are generally isoclinal, so that their limbs as measured in the field are parallel to the sheet dip. In the Sokumfjell Fold this is not the case. Therefore, the schist and quartzite bands within the marbles, and particularly the relations of their outcrop trends to variations in topography, are most valuable in making unambigous, if approximate, determinations of the sheet dip. In addition the schistosity in the pelitic schist bands is generally approximately parallel to the sheet dip.

Three main pelitic schist bands occur, which from the core of the fold outwards are the Sandvatn, Nævervatn and Lille Sokumvatn schists. The first two can be followed round the nortward closure of the fold, and occur in both limbs. The Lille Sokumvatn Schist, and the Navnløsvatn Quartzites between it and the Gneissic Group, occur only on the western side of the fold.

The sheet dip of the Sandvatn Schist is steep to the west in both limbs near Sandvatn, but northwards it becomes close to vertical in both limbs, which meet in a 'V' shaped outcrop. This implies that the axial plane of the fold must also be close to vertical and that the plunge must be steep. The western limb of the Nævervatn Schist shows a very steep dip to the west. The eastern limb also dips very steeply to the west near the fold closure, but southwards it changes gradually through the vertical so that it dips about 60° eastwards where it crosses Sokumaa. Thus an east—west section through the Nævervatn

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[1] It may be noted in passing that no structures which could be ascribed to the process of emplacement occur in the meta-sediments in association with these or any other granitic intrusions in the area.

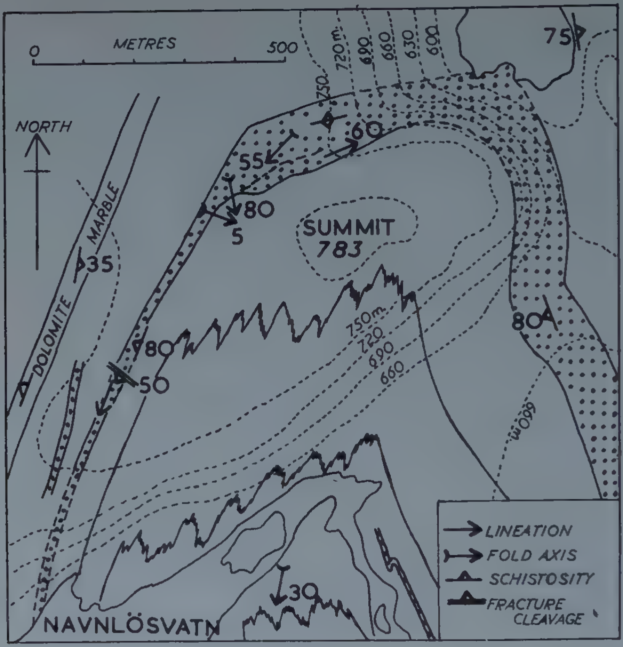

Fig. 4. Closure of the Sokumfjell Fold in the Nævervatn Schist. (Structure in marbles shown diagrammatically).

Schist indicates an antiformal structure and the northward closure would imply a northward plunge. However the relations at the 'U' shaped closure on summit 783 show that the plunge must actually be nearly vertical (fig. 4). The inner contact of the Nævervatn Schist runs up the steep southern slope of the hill without deviation in trend before turning west on the relatively flat top. The outer contact of the Schist stays on lower ground and runs to the S.W. corner of Nævervatn. It then turns west before climbing obliquely up the northern slope of the hill, again without change of trend. Thus the inner and outer contacts remain equidistant and parallel while the change of strike round the fold closure takes place, in spite of the different topographic relations. It follows that the sheet dip must be nearly vertical right round the fold closure, and the plunge must also be nearly vertical. This conclusion is confirmed by the attitude of the schistosity and bedding in the schist, which remain parallel to the contacts and are nearly vertical right round the fold closure. The local minor fold axes and lineations have various attitudes and are not homo-axial with the major structure.

It has already been mentioned that a similar plunge is indicated by the closure in the Sandvatn Schist. Further, there is a schist band, only a few metres thick, in the marbles between the two limbs of the Sandvatn Schist, which displays, on a smaller scale, similar U shaped fold closures to that of the Nævervatn Schist. Again the bedding and schistosity are nearly vertical all round the closures, though minor warps have various trends. In addition, fractures parallel to the vertical axial-plane are emphasised by thin granitic veins.

Therefore it is confidently inferred that the Sokumfjell Fold has a very steep plunge in a nearly vertical axial plane all the way from Sokumvatn to Nævervatn.

Further south, comparable evidence is not available to determine the plunge of the fold. The geometry is much less clear since the schist bands are interrupted or disappear completely. The western limb of the Nævervatn Schist thins and disappears some 600 metres north of Sokumvatn and both limbs of the Sandvatn Schist thin and disappear south of Sandvatn. The eastern limb of the Nævervatn Schist can be traced over Sokumfjell and several kilometres further south, until it finally disappears on HILL 830, east of Urfjell (fig. 5). The schist dips east at the northern end of Sokumfjell, but, coming south, it becomes vertical and then dips to the west at the southern end of Sokumfjell. The strike is also about N—S, and the dip steep to the west, on Hill 830; but between there and Sokumfjell the outcrop is 'S' shaped and the sheet dip varies. Two E—W striking portions have southerly dips, and they link together a portion which strikes N.N.W.—S.S.E. and dips steeply E.N.E. These relations suggest that the schist has been folded about axes plunging in various directions between S.E. and S.W. (fig. 5) and the linear structures and Beta maxima of fig. 3 *l* accord with this inference. As mentioned previously these structures are apparently related in origin to the Urfjell 'bulge'.

Other important schist outcrops occur east of Urfjell and south-east of Staburfjell. A dolomite marble band occurs on the north-western side of the latter outcrop. This band is apparently identical with the Bratfjell Dolomite which occurs outside the Nævervatn Schist

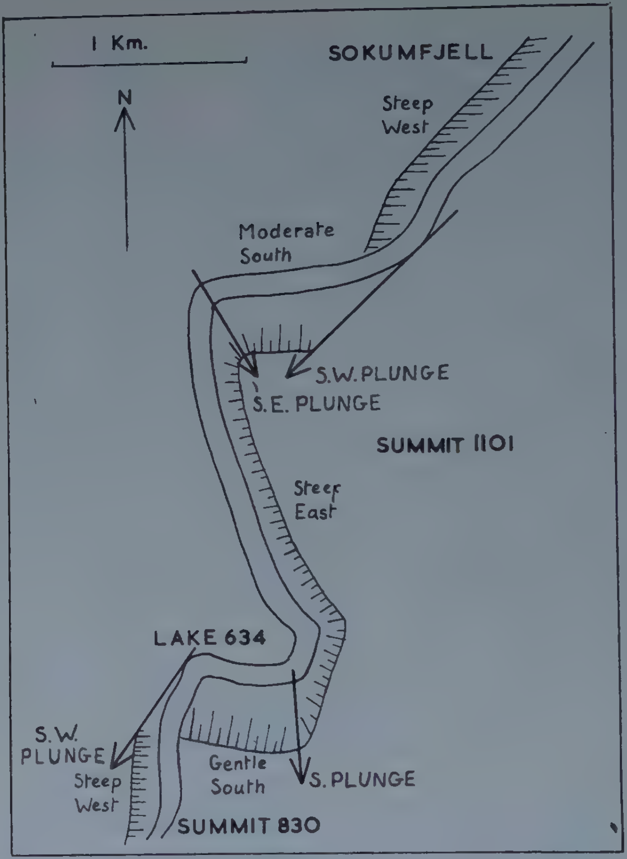

Fig. 5. Variations of sheet dip of the Nævervatn Schist.

in both limbs of the Sokumfjell Fold further north. No other dolomitic horizons have been found in the area and it therefore seems almost certain that the schist outcrops east of Urfjell and south-east of Staburfjell are to be correlated with the Nævervatn Schist. The sheet dip in both outcrops is under the Gneissic Group, steeper in the north-east and gentler in the south-west: that is, the dips conform with those of the Staburfjell Slide Zone in the Sokumvatn Synform.

Therefore, the axial plane trace of the Sokumfjell Fold can be carried southwards from Sandvatn between the two limbs of the Nævervatn Schist east of Urfjell, where the axial plane continues to dip to the west. The trend of the bedding in the intervening marbles shows that the axial plane trace then turns south-west and eventually westward to converge with the Nævervatn Schist south of Staburfjell (fig. 11). At the same time the axial plane dip changes from westerly to northerly.

Thus the axial plane trace of the Sokumfjell Fold swings round the Sokumvatn Synform, and its variation in dip also suggests conformity with that synform. When this is added to the fact that the fold is isoclinal and has a nearly vertical plunge at its northern end, it becomes evident that the Sokumfjell Fold cannot be an antiform complementary to the Sokumvatn Synform. Its geometry is readily explained however, if it is assumed to be an earlier overfold, like the Bjørnhatten and Harefjell Folds, which has been refolded by the Sokumvatn Synform.

The western limb of the Sokumvatn Fold is considerably thinner than the eastern limb. Thus the Bratfjell dolomite, the Nævervatn Schist, and the marbles which separate them, are very much thicker in the eastern limb than in the western limb where they are eliminated altogether towards the south. The variations are readily related to the tectonic position, but it cannot be proved, of course, that the thickness variation between the two limbs is not partly original, and in fact there is, apparently, a difference in the present lithology. The western limb contains a considerably higher proportion of schist and impure quartzite bands and of quartzo-feldspathic laminae in the marbles, than does the eastern limb. This can be explained by the tectonic hypothesis if it be assumed that the purer marble was preferentially squeezed out towards the crest of the fold during the thinning of the limb. However, the writer would not exclude the view that the variations in thickness due to tectonic effects may be superimposed on considerable original variations, which may have controlled the early development of the structure.

East of Urfjell the two limbs of the Nævervatn Schist are appreciably closer together than further north, and the Sandvatn Schist is absent from between them. This cannot be ascribed to a southward closure of the Sokumvatn Fold. The evidence continues to indicate a northward closure, and it is inferred that the core of the Fold to the north has been separated from its roots during the deformation (fig. 11).

The rocks between the Lille Sokumvatn Schist and the Gneissic Group have not been discussed in this section so far, since it seems that they may not belong in the western limb of the Sokumvatn Fold, although they conform to it. Several lines of evidence support this hypothesis, viz.: —

(1) When the Lille Sokumvatn Schist is followed north-eastwards beyond the present area, it shows no sign of turning southwards round the Sokumfjell Fold. In fact it begins to turn northward parallel to the contact between the Marble and Gneissic groups, before it disappears under ice.

(2) The Sandvatn and Nævervatn Schists generally form regular parallel-sided bands, but the Lille Sokumvatn Schist is irregular and deeply implicated with the adjacent marbles. This may indicate close proximity to a major slide which could separate the Nævervatn Schist from the Sokumfjell Fold.

(3) East of the Lille Sokumvatn Schist, the marbles of the Sokumfjell Fold display minor folds which are almost exclusively dextral[1] in sense. West of the Lille Sokumvatn Schist, on the other hand, the marbles normally show both dextral and sinistral folds in close association, and often the sinistral folds are most prominent.

(4) Two quartzite bands (the Navnløsvatn Quartzites) occur nearly parallel to each other between Navnløsvatn and Lille Sokumvatn, but north-north-eastwards from Lille Sokumvatn they diverge rapidly, the eastern band running parallel to the Lille Sokumvatn Schist, the western band parallel to the Staburfjell Slide Zone. The thickness of marble between them, measured normal to their sheet dip, increases from about 100 metres to 500 metres over a horizontal distance of only 1600 metres. There is no corresponding thickening in the marbles to east and west. It seems possible that the two quartzite bands belong to the same horizon, and that the marble which thickens so rapidly is in fact in the core of a fold complementary to the Sokumfjell Fold. This fold would be separated from the Sokumfjell Fold by the suggested slide east of the Lille Sokumvatn Schist (fig. 11) and further sliding would be necessary to account for the absence of a fold closure in the Navnløsvatn Quartzite.

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[1] Dextral and sinistral folds in the present context are those whose traces on a horizontal surface indicate dextral and sinistral rotations respectively.

Minor Structures in the meta-sediments.

The minor structures in the Sokumfjell Fold provide a marked contrast to those described from the western area, and they are much more complicated. They vary in style and attitude according both to the lithological type and to the position in the major structure.

Minor folds in the marbles are easily recognised by the numerous close set quartzo-feldspathic laminae, which mark the bedding. These laminae may owe their origin in part to mechanical differentiation of originally more homogeneous material but, if so, it appears that the differentiation has taken place parallel to the original bedding. The laminae are parallel to the undoubted sedimentary intercalations of pelitic schist and impure quartzite in the marbles.

The minor folds are extremely abundant in the axial region of the major structure. There, the plunges are extremely variable and may be either steep or gentle to the north or south. It seems reasonable to infer, in accordance with the interpretation of the large scale geometry, that the folds originally had steep plunges parallel to the plunge of the major structure, and that these have been subsequently modified by later movements. The data from the core of the Sokumfjell Fold north of Sokumvatn support this interpretation (fig. 3 k). It appears that the two limbs of the fold have been refolded about south or south-westerly plunging symmetrological axes (β^2 and β^3, which are maxima of bedding intersections from β diagrams) but that the axis of the major structure (β^1) is nearly vertical.

On the straight limbs of the major fold the minor folds in the marbles are much more consistent and statistically the folding may be nearly cylindroidal (fig. 3j). These folds have been studied particularly in the western limb of the Sokumfjell Fold, and especially in the Marmor Valley, which runs into the N.E. corner of Sokumvatn. Numerous folds occur with plunges of about 25° to S.S.W. Similar folds in more competent bands within the marbles often have steeper plunges. The folds have straight long limbs dipping steeply west, and highly puckered flat short limbs. The axial planes dip steeply west, nearly parallel to the long limbs (fig. 8a and b).

If these folds were supposed to be related to the formation of the Sokumfjell Fold as congruous drag folds, then they would suggest

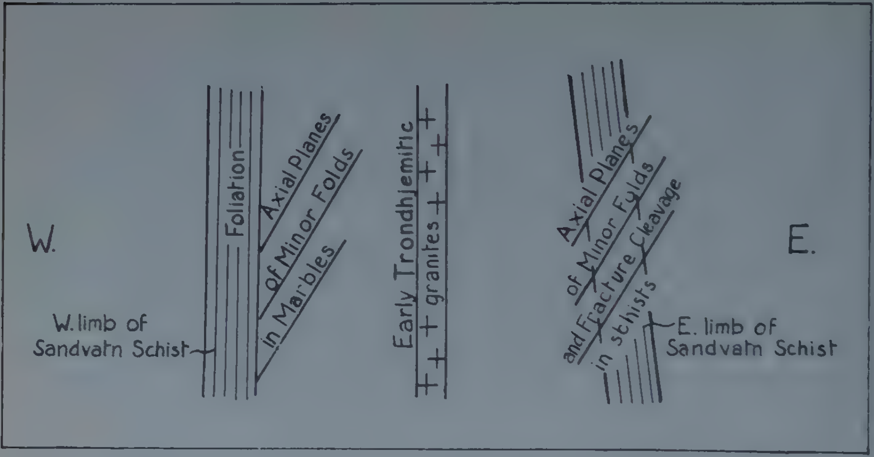

W.
W. limb of Sandvatn Schist
Foliation
Axial Planes of Minor Folds in Marbles
Early Trondhjemitic granites
Axial Planes of Minor Folds and Fracture Cleavage in schists
E. limb of Sandvatn Schist
E.

Fig. 6. Diagrammatic section across the Sokumfjell Fold.

that the latter is a syncline plunging to the S.S.W. In fact, it is known that the plunge of the Sokumfjell Fold is nearly vertical, but even without this the above explanation could be discounted, since, in a section drawn normal to the S.S.W. plunging minor fold axes, the Sokumfjell Fold is antiformal, not synformal. Further, the axial plane of the Sokumfjell Fold must be nearly vertical, between the two limbs of the Sandvatn Schist. Yet the axial planes of the minor folds in the marbles and fracture cleavage in the schists cut across both limbs in the same direction (fig. 6).

It can hardly be doubted therefore that these minor folds are not coeval with the Sokumfjell Fold, but must be regarded as superimposed upon it. Their axial trend clearly suggests that they should be correlated with the formation of the Sokumvatn Synform. Even so their sense is not appropriate to congruous drag-folds in relation to the synform. This is not readily explained, but similar relations elsewhere have been ascribed to shear folding (FAIRBAIRN 1949, pp. 160—161) or as 'flowage' folding (BAIN 1931).

In the pelitic schist bands the most prominent structure is the schistosity, which is approximately parallel to the contacts both in the fold limbs and round the fold closure. Thus on the largest scale the schistosity approximates to the bedding, and is certainly not axial plane to the major structure. In detail it can often be seen that the schistosity is not strictly parallel to the bedding, but there is little evidence that it is then to be regarded as axial plane to isoclinal folds, as it was in the Meløy schists. In only a few cases have minor folds definitely belonging to the first folding been distinguished. Thus the schistosity must be regarded essentially as a bedding schistosity produced by bedding plane slip during the formation of the Sokumfjell Fold. This schistosity must have suffered considerable modification during the formation of the Sokumvatn Synform in the second folding since lineations (due to dimensional orientation of mica, and sometimes of sillimanite and tourmaline) with gentle southerly plunges are commonly developed.

Small scale flexure folding of the schistosity occurs sporadically (fig. 8c-f). The sense of these folds varies, but they more commonly indicate movement appropriate to normal drag folds in the Sokumvatn Synform. They generally have gentle southerly plunges and clearly also are to be correlated with the second folding; they have been superimposed on the earlier ubiquitous schistosity of the first folding (cf. WHITE and JAHNS 1950, pp. 196—200). Occasionally a new schistosity develops parallel to the axial planes. In particular lithologies an axial plane fracture cleavage is developed rather than a new schistosity (cf. BROUGHTON 1946, p. 7 and p. 11). This fracture cleavage is of the 'false cleavage' type (LEITH 1905, p. 120). The schistosity is thrown into small puckers, defined by generally unbent biotite flakes. The closely spaced shear planes are simply the axial planes of these puckers, and there is no rupture along them. A sillimanite lineation is common parallel to the intersection of the cleavage with the schistosity.

The mica fabric of the schists shows a partial girdle pattern normal to the macroscopic lineations of the second folding (fig. 9j, l, m). The schistosity as defined by the main biotite maximum may be parallel to the limbs (fig. 9 m) or to the axial plane (fig. 9 l) of the minor folds mentioned above. In one case where the 'bedding' could be distinguished (fig. 9 l), the axis of the mica girdle was clearly not parallel to the 'bedding' (S_1) and made a large angle with the minor fold axis (F), although the new schistosity (S_3) was clearly axial plane to the minor fold. Thus it may be inferred that the girdle axis represents an axis of internal rotation of an earlier schistosity (S_2?) which was

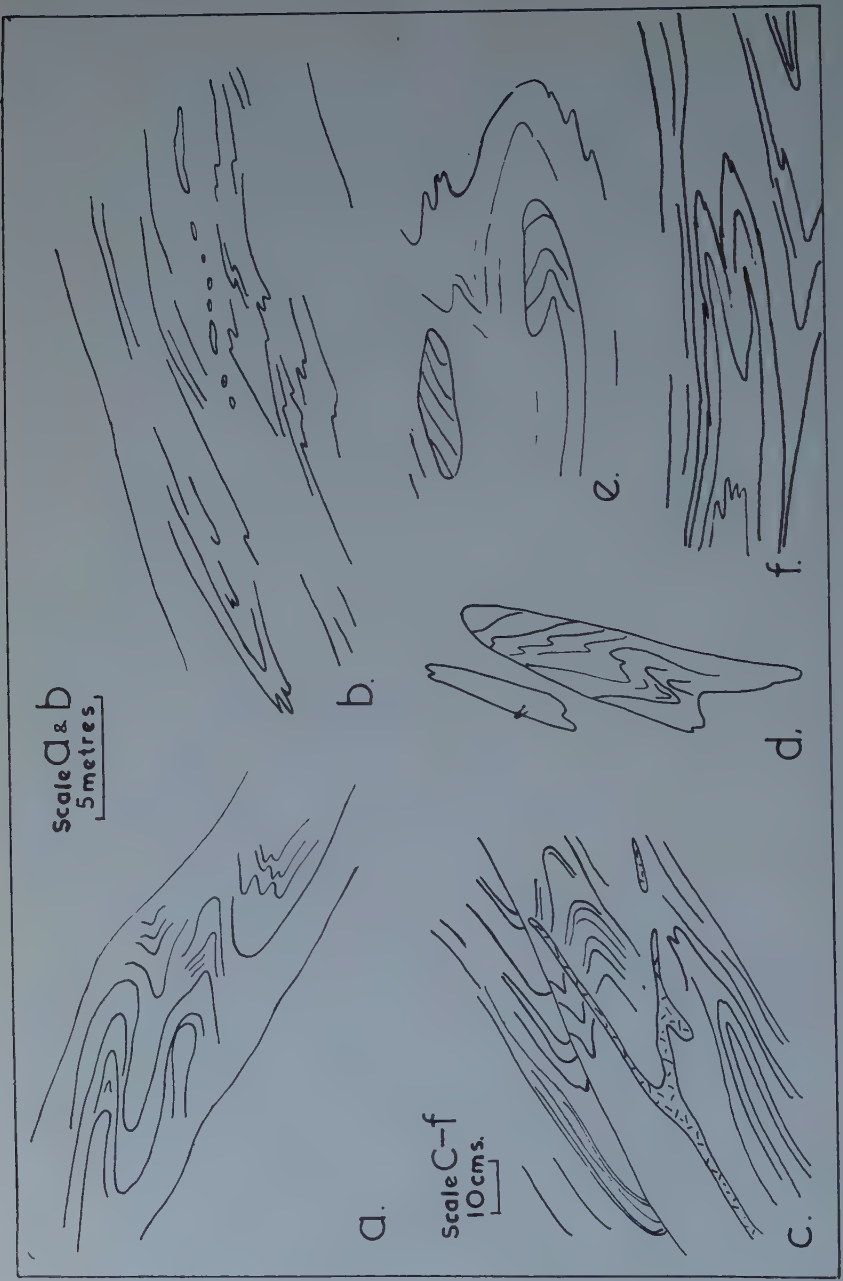


Fig. 7. Minor structures – first folding.

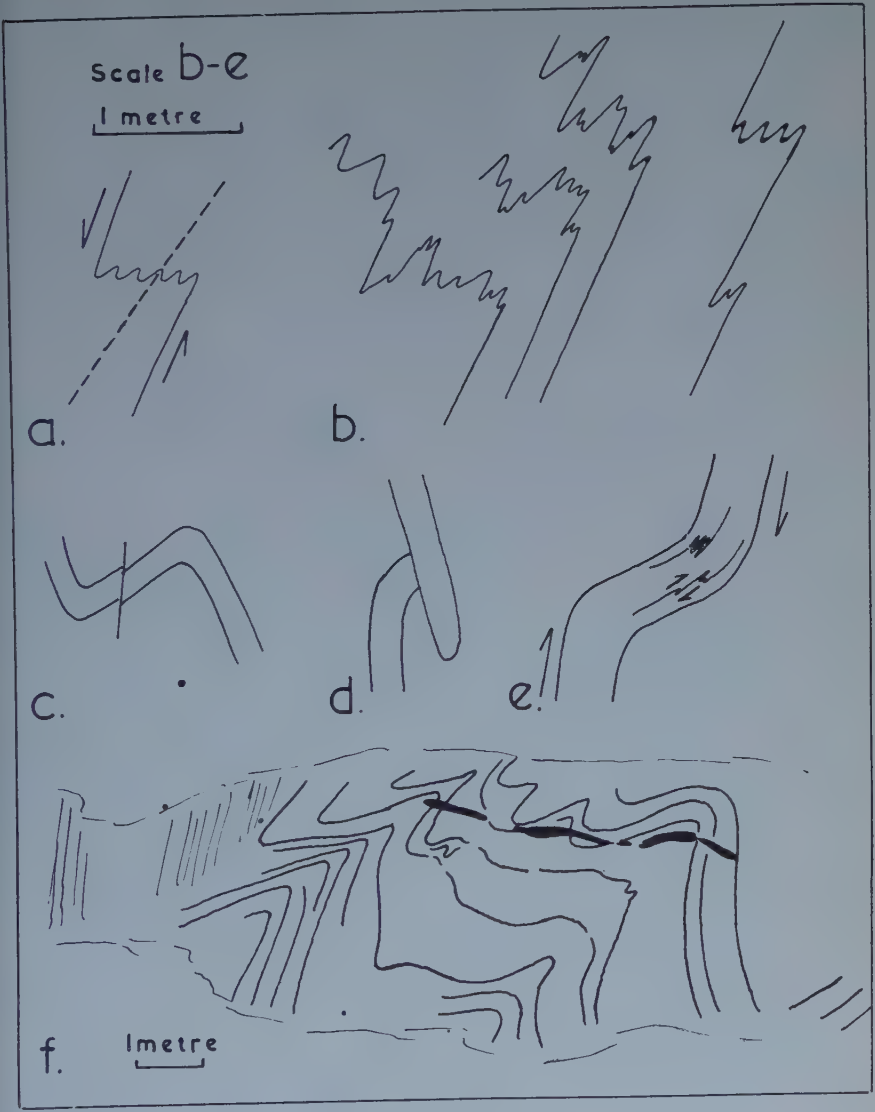


Fig. 8. Minor structures — second folding.

itself not parallel to the bedding. This supports the inference made above from the relationship to the major structures that these lineations must have been produced by the modification of an earlier schistosity.

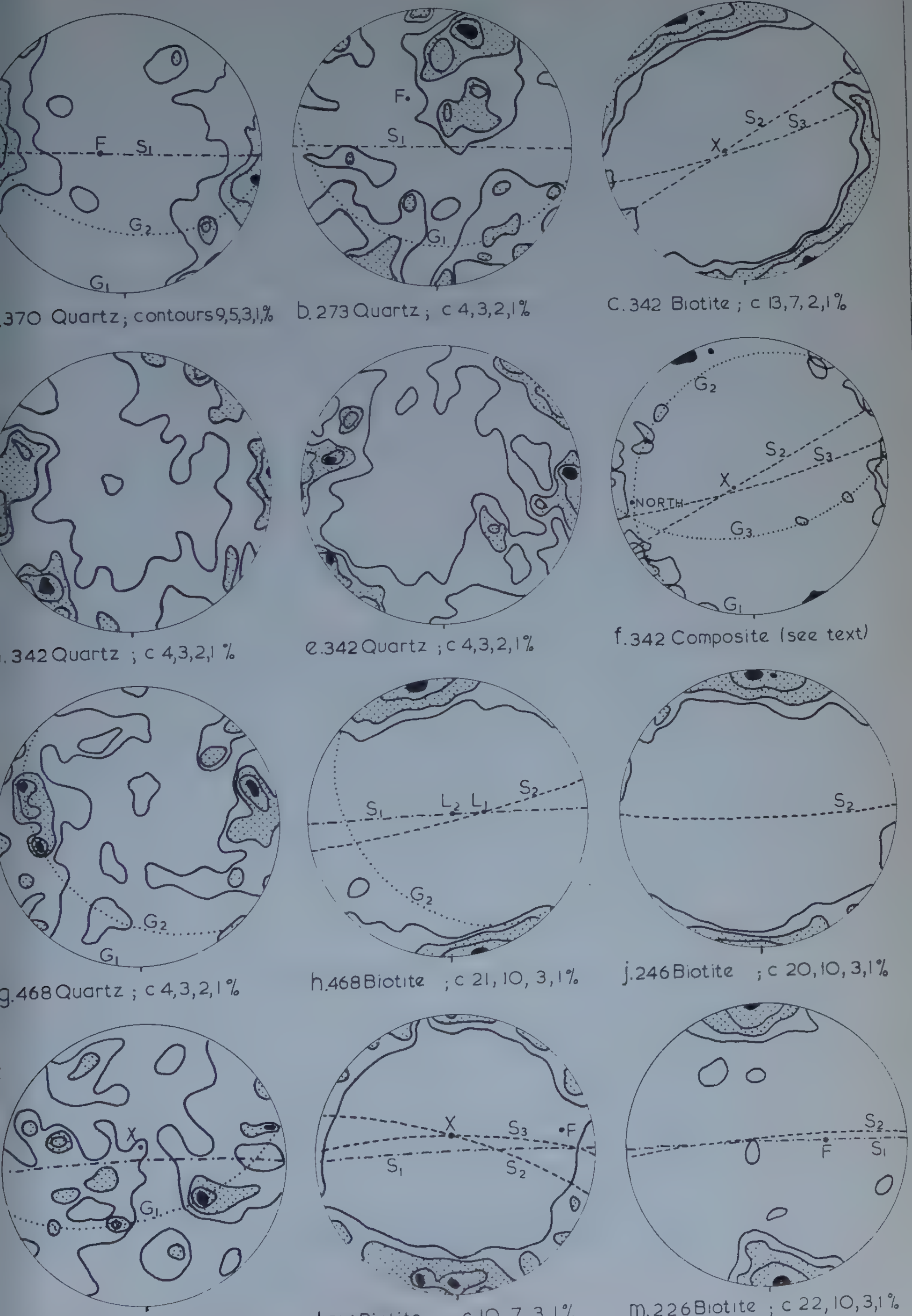
The quartz fabrics of the schists are virtually random; where a faint girdle can be distinguished it is not homotactic with the mica girdle (fig. 9 k & l). Thus it appears that, in the schists, the quartz has not been reorientated as readily as the mica during the second folding. This is probably due to the fact that the schists are highly feldspathic, containing only about 10 % quartz, so that the quartz reorientation has been interfered with by the feldspar. As noted below, pure quartzites have developed a new girdle orientation during the second folding.

The Navnløsvatn Quartzite bands show alternations of massive, pure quartzite with laminated impure quartzite, the former having commonly been boudined. The deformation accompanying the boudinage must have been relatively plastic, since the individual boudins are drawn out to tapering terminations. Further, the axes have very steep plunges, and it therefore seems that the boudinage can be correlated with the first folding. Some tight dextral minor folds with very steep plunges probably also belong to the first folding, but in addition there are open sinuous flexures of the quartzite bedding, generally on steep N.W. plunging axes, which most probably belong to the second folding. Rarely, a mineral lineation of slickenside type, with a plunge roughly down dip, has been seen on bedding surfaces, thus indicating late, mainly vertical, relative movements between adjacent beds in contrast to the boudinage and most minor folds.

Examination of the fabric of two quartzites showed that it has been adjusted to the second folding. A strong preferred orientation in one diagram (fig. 9 a) shows a 9 % maximum in a partial girdle (G_1) which is roughly normal to the second fold axes (F) as shown by the adjacent marbles. A similar girdle (G_1) may be recognised in a second quartzite (fig. 9 b) but in this case it has been almost obliterated by post-tectonic crystallisation associated with the granitisation of the region.

Fig. 9.

Fig. 9. Micro-fabric diagrams. Specimen localities shown on map to fig. 3. Qua
diagrams are of 200 optic axes; mica diagrams are of 200 poles to cleava



.370 Quartz ; contours 9,5,3,1,%

b. 273 Quartz ; c 4,3,2,1%

c. 342 Biotite ; c 13,7,2,1%

. 342 Quartz ; c 4,3,2,1 %

e. 342 Quartz ; c 4,3,2,1%

f. 342 Composite (see text)

g. 468 Quartz ; c 4,3,2,1%

h. 468 Biotite ; c 21, 10, 3, 1%

j. 246 Biotite ; c 20, 10, 3, 1%

k. 214 Quartz ; c 4, 3, 2, 1 %

l. 214 Biotite ; c 10, 7, 3, 1%

m. 226 Biotite ; c 22, 10, 3, 1 %

Kinematics of the Sokumfjell Fold.

The geometry of the Sokumfjell Fold as described above provides clear evidence of the nature of the large scale movements which have produced it. It has been shown that the plunge of the structure north of Sokumvatn is nearly vertical, on that the map provides a section virtually normal to the first fold axis.

The schist bands are continuous round the fold closure, and they display a schistosity which follows the bedding round the closure. There was no shear across the bedding in the axial plane direction, and the schist bands have clearly attained their present form by flexure accompanied by bedding plane slip.

The marbles between the schist bands however show great thinning of the fold limbs and great thickening of the axial region. If the second folding is ignored, the bedding in the limbs is straight and parallel to the schist bands while the axial region shows concertina folding (fig. 4). Clearly the thickening in the axial region has been produced by movements, from the limbs towards the crest, of the outer layers over the inner. Again, whatever the detailed mechanism of plastic flow in the marbles, the movements have taken place essentially by bedding plane slip.

The demonstration of the Sokumfjell Fold as a flexural slip fold is of particular interest in view of the kinematics which have been suggested for major fold structures in carbonate rocks in other orogenic zones (Cloos 1947, Bain 1931). Axial plane shear movements could not account for the observed geometry of the Sokumfjell Fold, and there is no evidence of such movements.

Structural relations of intrusive bodies.

The Sokumfjell Marbles are intruded by an abundant suite of intrusive rocks which vary from trondhjemitic to true granitic composition. There are also some thin amphibolites which probably represent original dolerite intrusions.

The amphibolites occur in the marbles as small masses extending a few metres along the strike and one or two metres across it. It is usual to find a number of these masses on the same general strike, separated often by some tens of metres. The masses are regarded as boudins, with very steep axes. The schistosity and foliation within the boudins is planar and roughly parallel to the strike of the enclosing marbles and there is therefore no suggestion that individual boundins represent an originally greater strike length of amphibolite which has been 'rolled up'. The separation of the boudins therefore indicates enormous stretching parallel to the present strike of the marbles, or normal to the now vertical axis of the Sokumfjell Fold.

Amongst the granitic rocks, the potassic microcline-granites usually show little or no deformation. They often follow the bedding in the folded marbles, though they may also form an irregular network. There is also a very late suite of narrow, planar, microcline granites which cut right across the bedding with W.N.W.—E.S.E. strike. Definite boudinage is rarely seen in the potash granites, though some 'pinch and swell' structures may be due to tectonic effects.

Attention is here focussed on a suite of trondhjemitic dykes in the axial region of the Sokumfjell Fold as seen in the Marmor Valley. These are parallel sided intrusions often about one or two metres thick, with a constant N—S strike cutting across the N.N.E.—S.S.W. strike of the marbles. Their dip always appears to be very steep, greater than that of the enclosing marbles. Thus they agree in attitude with the axial plane of the Sokumfjell Fold. The dykes have invariably suffered boudinage and the individual boudins are usually three or four times as long as they are broad, measured normal to their axes. The ratio of the lengths of the inter-boudin gaps to the length of the boudins is perhaps about 1:1. The ends of the boudins may be clear fractures or they may be drawn out to pointed terminations (fig. 10). It is difficult to determine the plunges accurately since the intrusions usually weather out only a few centimetres above the surrounding marbles but it is clear that the plunge is generally very steep. Thus it is roughly parallel to the inferred axis of the Sokumfjell Fold and shows no agreement with the minor folds in the marbles which have plunges of about 25° S.S.W.

The marbles at the margins of the dykes have invariably suffered drag folding against the dyke walls, and these folds always indicate the operation of a dextral movement couple in the horizontal plane (fig. 10). The plunges of these folds vary from very steep to the north to steep or moderate to the south. Away from the dyke walls there is a rapid but continuous change to the gentle S.S.W. plunges previous described.

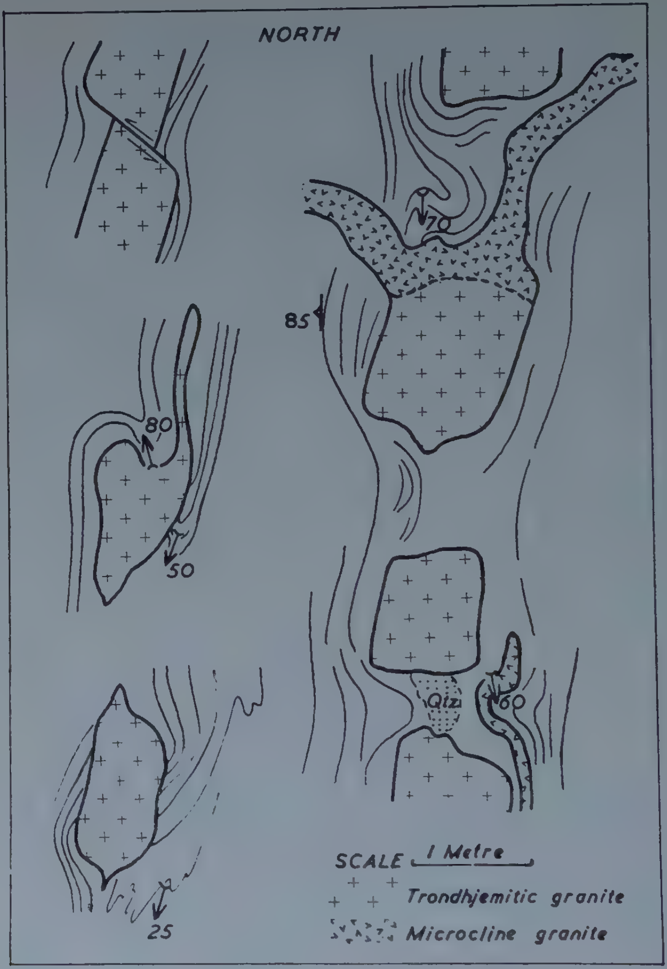


Fig. 10. Boudinage of discordant granites in marble.

In considering the origin of the boudinage with respect to the two main periods of folding, it must be recognised that the emplacement of the dykes, the separation of the boudins and the drag folds on the dyke walls are three distinct phenomena which may be of different ages. Also the boudinage is unusual in that it is produced in discordant intrusions, so that the rules which apply to the development of boudinage in conformable competent layers may not be applicable. In particular, boudin axes need not necessarily be expected to form parallel to the fold axes, unless the intersection of the discordant dyke with the bedding is itself parallel to the fold axes. If the intersection on the bedding is otherwise, the directions of maximum shear or tension in the dykes are unlikely to be parallel to the fold axes.

The dykes in the Marmor Valley are not exceptional in following the axial plane direction of the Sokumvatn Fold, since granites parallel to the axial planes of the first folds are common in the western area where first fold structures are dominant. Everywhere these granites are the earliest in the sequence, and the structural relationship shows clearly that they are essentially later than the first folding. The boudin axes in the Marmor Valley dykes are however roughly parallel to the first fold axes, and they may suggest therefore that the boudinage was produced in the later stages of the development of the Sokumfjell fold during the first folding. A continuation of rotation and extension of the marbles between the dykes could lead to extension of the dykes parallel to the axial plane. The movement involved would obviously be very considerable, and it is difficult to envisage its taking place without a displacement of the boudins from their initial planar arrangement. Further it may be questioned whether the emplacement of the dykes could have taken place while the maximum principal stress continued to act across the axial plane of the Sokumfjell Fold.

Alternatively it may be supposed that the separation of the boudins was brought about during the second folding. It has already been mentioned that boudinage in similar granites in the western area must be ascribed to the second folding, but the difficulty in the Marmor Valley is that the boudin axes are roughly parallel to the first fold axes. This may not be an insuperable objection, since the boudin axes may be controlled by the intersection of the discordant dykes with the bedding. The directions of greatest tension in the dykes during the second folding may have been roughly parallel to this intersection, although the minor fold axes have a very different plunge. In any case the drag folds on the dyke walls must certainly be ascribed to the second folding. They show a compromise in attitude between the original intersection of the bedding with the dykes and the minor fold axes of the second folding. Their sense is appropriate to the movements which produced the latter minor folds, but it could not be correlated with shear parallel to the axial plane in the first folding, since this would be in the opposite sense.

It may be concluded therefore that

(1) the dykes were emplaced after the first folding.
(2) the boudinage may have been produced in the later stages of the first folding or during the second folding.
(3) the drag folds on the dyke walls were produced during the second folding.

Finally it may be noted that the boudinage was certainly earlier than the latest potash granites, since these may pass through the inter-boudin gaps. Occasionally however these granites have also suffered boudinage.

THE FISKEVATN FOLD

The axial plane trace of the Fiskevatn Fold shows a slight swing round the Sokumvatn Synform, like the Sokumfjell Fold. But in contrast to the latter, the limbs of the Fiskevatn Fold dip away from the Sokumvatn Synform.

South of Staburfjell, the Nævervatn Schist dips about 40° northwards under the Gneissic Group, but two kilometres to the south the Fiskevatn Schist is vertical or dips steeply south (section EE′ fig. 2). Again, on Urfjell, the Nævervatn Schist dips about 75° N.W., but two kilometres to the S.S.E. the Fiskevatn Schist dips at less than 40° to S.E. Thus both these lines of section suggest that the sheet dip is controlled by folding with a N.E. axis. Either there is an antiformal structure or the Fiskevatn Schist is inverted relative to the Nævervatn Schist. In the intervening marbles the minor folding makes the sheet dip difficult to determine, but the observations suggest that the greater part of the change in sheet dip, between the Nævervatn and Fiskevatn Schists, takes place suddenly.

North-east of Lake 520, the Fiskevatn Fold is isoclinal. The Fiskevatn Schist dips to the south-east at not more than 40° in both limbs. South-west of Lake 520 however, in the region of Sundvatn, the fold is more open with an angle of about 60° between the limbs. The north-west limb of the Fiskevatn Schist is here nearly vertical, but the marbles of the south-east limb dip only about 30° to S.E. In between, the axial region of the fold is shown by 'concertina' folding in the marbles and by large scale folding in a minor schist band. These structures indicate that the axial plane of the major structure must here have a S.S.E. dip of about 65°—70°.

North of Sundvatn the minor fold axes and lineations generally plunge gently E.N.E. The structures agree both in style and trend with those of the first overfolding in the western area, and it is therefore inferred that the Fiskevatn Fold is also an overfold produced during the first folding. As in the western area, a few later flexure folds occur on axes trending just west of north and these are correlated with the second folding. North-east of Lake 520 there is little change in the trend of the axial plane trace, and it is supposed that the fold continues to plunge in an easterly direction. However, the fold has become overturned and isoclinal and the change of sheet dip may not be an original effect of the first folding but may be due to refolding on N.E. axes (section EE' fig. 2). The dominant minor folds in this region plunge S.E. and have a style similar to that of the minor folds in the Sokumvatn Fold. They are therefore correlated with the second folding. Their trend is not far removed from that seen on the E. limb of the Sokumfjell Fold but they now plunge down the sheet dip. Thus if the sheet dip owes its present attitude to second folding on N.E. axes these minor folds are to be regarded as cross folds to the local main direction of second folding. The β and πs diagram for the Fiskevatn Fold (fig. 3 m) gives some support to this hypothesis. Each limb appears to have been refolded about new axes (β^2 and β^3) but the major structure has an easterly plunge (β^1) (cf. fig. 3 k). In the area to the south it can be shown much more certainly that the second folding has involved two main directions at right angles (B. J. Walton, personal communication).

It is not possible to make any certain correlation between the rocks of the Fiskevatn and Sokumfjell Folds, and it is therefore difficult to assess the significance of the apparent antiformal attitude of the sheet dip between the two structures. Since the change of sheet dip appears to take place suddenly, it is possible that the two folds are separated by a major slide, which would also account for the failure of the eastern limb of the Sokumfjell Fold to continue southwards (fig. 11). Other interpretations are also possible, but discussion must be deferred until the structure of the ground further south is fully known. In the present context, the chief importance of the Fiskevatn Fold is in providing further evidence of the first overfolding on E.N.E.—W.S.W. axes, and in showing that the Sokumvatn Synform does not extend south of Fellvatn.

Relationship of major and minor structures

It has already been emphasised that the axial method of tectonic analysis does not apply to the Sokumvatn Area as a whole. The plot of linear structures (fig. 3a) indicates the range of variation of trends and plunges in the area, but it is heavily weighted to the abnormal trends, and fails to give proper emphasis to the dominant trends.[1]

These dominant trends may be summarised as follows (see also fig. 11): —

(1) The western half of the area, corresponding to the western limb of the Sokumvatn Synform, has dominant E.N.E.—W.S.W. trending linear structures of the first period of folding. N.N.W.—S.S.E. trends of the second folding are subordinate. (fig. 3h).

(2) The Sokumfjell Fold shows dominant linear structures plunging gently S.S.W. in its western limb and plunging gently S.S.E. in its eastern limb. Aberrant trends occur throughout the quadrant between S.S.E. and W.S.W. and to some extent in the opposite quadrant (fig. 3d).

(3) The Fiskevatn Fold shows dominant E.N.E.—W.S.W. trending structures of the first folding east of Fellvatn but, further east, S.E. plunging structures of the second folding are dominant.

The dominance of second fold structures in the eastern limb of the Sokumvatn Synform, compared with a relatively minor developement of these structures in the western limb, must be ascribed to much more intensive deformation in the second folding, as might be expected in the steep limb.

The considerable variations in the attitudes of the minor linear structures can be explained by the operation of two factors:

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[1] Repeated measurements of the dominant trends were not made, partly because of the difficulties of making really accurate measurements in the marbles and partly because the process would have taken too much of the available time.

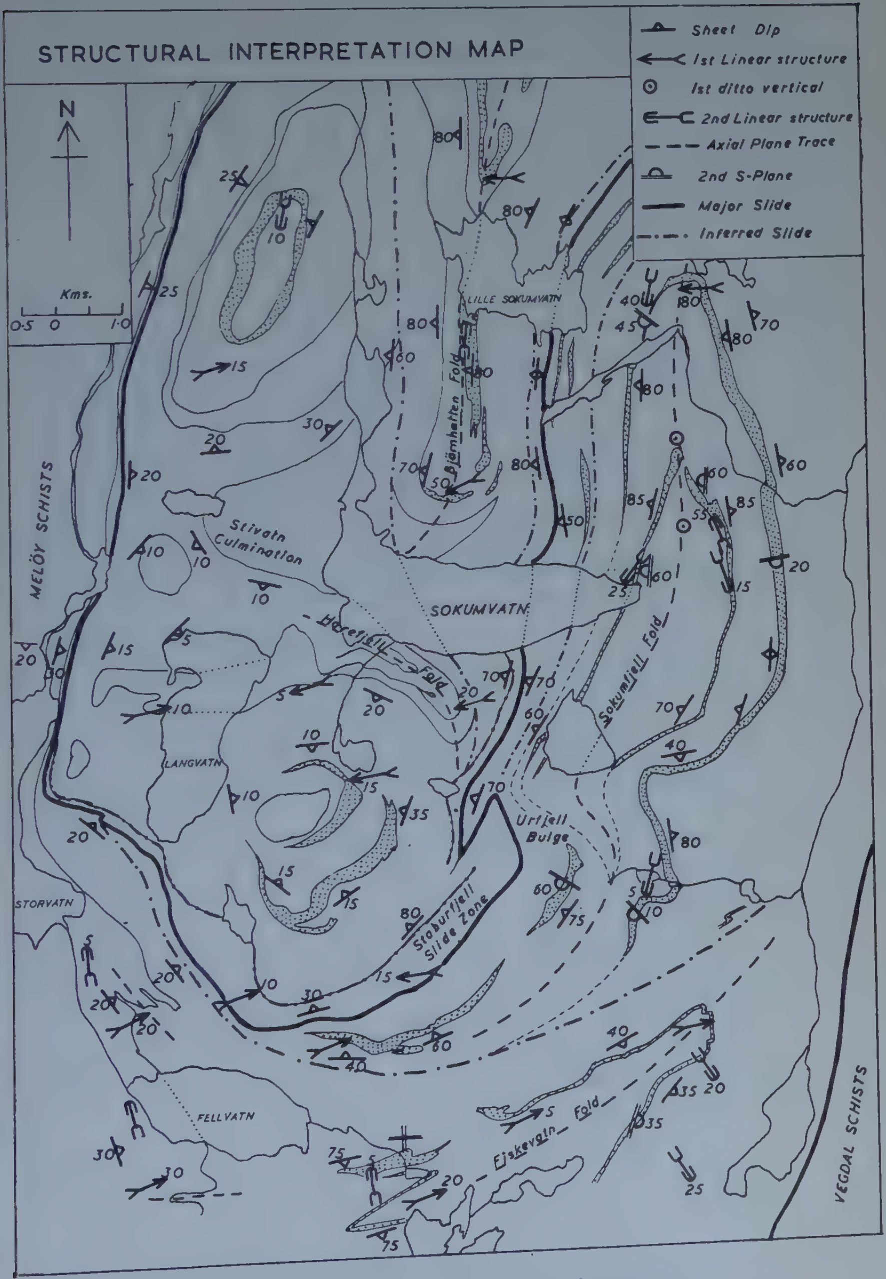

Fig. 11. Structural Interpretation map.

(1) External rotation of first fold structures on various possible axes during the second folding.

(2) Dispersion of axes of internal and external rotation of the second folding as a result of the variations in attitude of the pre-existing planar structures. This factor may operate directly, or by causing a local variation of the stress pattern during the second folding.

(1) The external rotation of first fold structures has been demonstrated on all scales. It has been shown that the Harefjell, Bjørnhatten and Sokumfjell folds are isoclinal folds whose plunges are to be explained by their refolding in the Sokumvatn Synform. Minor folds belonging to these major structures have suffered similar rotations. The change across the axis of the Sokumvatn Synform from gentle E.N.E. plunges in the western area to rather steep S.W. plunges in the Bjørnhatten Fold and finally to nearly vertical plunges in the Sokumvatn Fold may be ascribed primarily to this factor. Some scatter of the resultant plunges is to be expected for any given amount of rotation since the external rotations will take place on various axes, as noted below. Moreover, the external rotations are likely to be accompanied by various internal rotations as the intensity of the second folding increases. This was exemplified above by the development of new lineations and S-planes in the pelitic schists of the Sokumfjell Fold. The microfabrics of two specimens of pelitic schist (Nos. 341 & 468) from the Staburfjell Slide Zone are described here to illustrate the earlier stages of the modification of the first fold fabric.

The specimens were taken from either side of the axis of the Sokumvatn Synform (see map, fig. 3). At each locality intense isoclinal folds on E.N.E.—W.S.W. trending axes belonging to the first folding were observed, and no minor folds with N—S trending axes were seen. Each fabric shows approximately the expected 'ac' girdles normal to the megascopic first fold lineation; but a slight difference in orientation of the quartz and mica girdles is to be observed in the more westerly specimen (No. 342); and a much more marked difference in the more easterly specimen (No. 468) makes the quartz and mica fabrics definitely heterotactic.

Figs. 9c, d and e show biotite and quartz fabrics for specimen 342. In all, three quartz diagrams were prepared to test the reproducability of the results and it was found that, although the maxima shift in position (the fabric is not completely homogeneous) they are related to definite S-planes. The composite diagram (fig. 9f) shows the areas of greater than three per cent concentration from all three quartz diagrams as well as the main biotite maxima (black). The main S-plane (G_1) recognizable from the quartz fabric is roughly equatorial in the diagram and is slightly offset from the biotite girdle, and two weak subsidiary S-planes (G_2 and G_3) can also be recognised. These features can be ascribed to slight overprinting during the second folding, and the two subsidiary S-planes (G_2 and G_3) intersect with each other and with the subsidiary biotite S-plane (S_2) in a northerly direction which may represent the axis of the second folding.

Figs. 9g and h show biotite and quartz fabrics for specimen 468. The main biotite S-plane is oblique to the bedding and intersects the latter in L_1 parallel to the macroscopic lineation. A further linear structure (L_2) is indicated by slight rippling on the bedding surface; it has no obvious significance unless the biotite girdle is actually composite. The quartz maxima show a clear departure from the biotite girdle and suggest development of a new S-plane (G_2). The fabric is clearly hereotactic and triclinic and may be regarded as showing a more advanced stage of overprinting by the second folding than that of specimen 342.

Thus the tendency toward development of a heterotactic triclinic fabric can be ascribed to differential modification of the quartz and mica fabrics during the second folding. It has been noted already that the fabrics tend to remain heterotactic and triclinic even when the modification during the second folding has been more extreme, as in the pelitic schist bands in the Sokumfjell Fold.

(2) The dispersion of the axes of internal and external rotation of the second folding is a more complicated problem. Before the first folding it can be expected that the bedding had a fairly constant attitude, so that the axes of internal and external rotation of the first folding would be fairly constant and parallel to each other. Before the second folding on the other hand the pre-existing structure is much less simple. The stress-strain relations are likely to be complex, and the attitude of the pre-existing structures will considerably influence that of the resultant structures (e.g. RAMSAY 1958, WEISS and MCINTYRE 1957). Therefore differences of second fold trends are

to be expected in the Sokumvatn Area between the two major rock groups, which were structurally unconformable after the first folding, and between opposite limbs of major overfolds of the first folding.

The symmetrological axis of the Sokumvatn Synform, as seen in the Gneissic Group and Staburfjell Slide Zone, appears to be N—S, or slightly east of north. Some minor structures in the Gneissic Group share this trend, but the axes of minor structures in the Marble Group diverge from this in varying degrees. In the western area, the second fold axes are nearly horizontal and strike N.N.W.—S.S.E., approximately at right angles to the first fold axes. In the western limb of the Sokumfjell Fold the linear structures commonly plunge S.S.W.: in the eastern limb they plunge S.S.E. This is clearly the consequence of the steep attitude of the limbs of the Sokumfjell Fold and their divergence southward. As a consequence of this divergence, folds in opposite limbs may have a common axial plane direction and very different plunges. Thus in the Marmor Valley the strike of the marbles is N.N.E.—S.S.W. and the sheet dip nearly vertical. The minor folds, which have gentle southwesterly plunges, have axial planes dipping about 60° to the west. In the eastern limb of the Sandvatn Schist the sheet dip is again nearly vertical, and folds occur with similar axial plane direction, but the bedding strike is now N.N.W.—S.S.E. and the minor folds plunge at over 50° to N.W.

In addition it may be noted that different lithologies within the Sokumfjell Fold have reacted differently during the second folding. The marbles have readily developed second minor folds and the pelitic schists lineations with gentle plunges. But minor folds in more competent calc-silicate bands within the marbles often have steeper plunges than those of the marbles; and the thicker quartzite bands only develop warp structures with steep plunges.

It is not possible to say with certainty whether the variations described above are directly due to the variation in the attitudes of the S-planes before the second folding, or whether there was some local variation of the stress pattern during the second folding. This can best be tested by graphical statistical analysis of the attitude of shear planes (e.g. fracture cleavage) produced during the second folding (cf. WEISS and MCINTYRE 1957). If the stress pattern was uniform over the whole area, these planes in stereographic projection should have a common intersection corresponding to the intermediate principal stress direction. Any scatter of the intersections would be a measure of the variations in the stress pattern. In the Sokumvatn Area the data is inadequate to draw definite conclusions, but the scatter of the fracture cleavage plane intersections does suggest considerable variation of the stress pattern.

Age relations of the fold episodes

It is fundamental to the structural interpretation developed in the preceding sections that the folding which produced the Sokumvatn Synform is later than the major overfolds on E.N.E.—W.S.W. axes and has refolded them. In some other areas double fold systems have been separated into two fold sets of different ages (e.g. RAMSAY 1958), but in others again the two fold sets are believed to have been formed virtually simultaneously (e.g. KING and RAST 1955). In the Sokumvatn Area there is good evidence that the two fold episodes are distinctly separated in time. The second episode does indeed involve folding on two directions at right angles when a larger area is considered but only one direction is important in the Sokumvatn area.

STRUCTURAL STYLE

A marked difference in structural style between the first and second folding can be recognised both in the major and the minor structures which have been described. The main points of contrast are summarised here.

The major overfolds of the first folding were isoclinal and recumbent before the second folding. They are each confined to a single major lithological unit and important slides occur between the units. The Sokumvatn Synform, on the other hand is a relatively open structure with an inclined axial plane, and involves both the major lithological units. This is itself indicative of a less plastic deformation at a higher structural level.

The first minor folds in the pelitic schists are isoclinal with an axial plane schistosity. The associated quartz and mica fabrics are homotactic, showing an 'ac' girdle normal to the fold axes and lineation. The second minor folds however are generally open and the earlier schistosity is parallel to the fold form. A new axial-plane fracture cleavage or schistosity sometimes develops in the second folding. The associated quartz and mica fabrics tend to be heterotactic, only the mica girdle being clearly related to the second folding.

The first minor folds in the marbles are isoclinal like those in the schists, and both limbs of the folds are long (fig. 7f). The second minor folds however are open, and there is a marked contrast between straight long limbs and highly puckered short limbs (fig. 8a and b).

These characters all indicate that the second folding was much less plastic than the first. It is concluded, therefore, that the two fold sets were formed under distinctly different physical conditions. The latter cannot be due to difference in place, since the second fold structures are actually superimposed on the first. It follows that the different conditions imply a considerable difference in time between the two fold episodes.

GRANITE EMPLACEMENT

The early trondhjemitic granites, described from the Marmor Valley emphasise the difference in time between the two fold episodes. They were emplaced essentially after the first folding as planar intrusions parallel to the axial-plane of the Sokumfjell Fold. Subsequently they were boudined; and dragfolds were produced on their margins during the second folding.

The rocks of the Gneissic Group display various stages of metasomatism, so that the anorthite content of their plagioclase varies from over 40 % to less than 15 %. The rocks of the Marble Group show a sequence of granite veins from early trondhjemitic granites (with andesine) to late microcline rich granites (with acid oligoclase). Thus it may be inferred that during the granitisation the composition of the introduced material was becoming progressively more potassic, and the temperature was falling. Even the most potassic granites have occasionally suffered some deformation and it seems likely therefore that the second folding took place at a considerably lower temperature than the first folding which preceded the emplacement of the trondhjemitic granites.

MINERAL PARAGENESES

Ackermann's detailed studies of the mineral parageneses bear out the hypothesis of falling temperature during the granitisation (personal communication). The lowest temperature mineral to develop is talc in dolomitic marbles which had developed diopside at the height of the Regional Metamorphism. This talc is sometimes deformed by fracture cleavage. Thus it appears that deformation continued to a very late stage in the sequence of granitisation and mineral development. However, it seems unlikely that the second folding as a whole can have taken place so late in the sequence. In some pelitic schists sillimanite forms a lineation[1] parallel to the intersection of bedding and fracture cleavage, and clearly it has developed during or after the development of the second folds. Metasomatic sillimanite has been described (e.g. WATSON 1948) and its stability relations are still inadequately known, but present evidence suggests that its growth implies at least a moderately high temperature of metamorphism, whatever the pressure (e.g. CLARK et al., 1957).

It is clear that the three main lines of evidence cited above all support the contention that the two main fold episodes took place under different conditions and were clearly separated in time, though no stratigraphic dating can yet be offered. The second folding took place in a different stress field and probably at a lower temperature than the first folding. Possibly also the hydrostatic pressure was materially different. An uplift with a reduction of hydrostatic stress after the first folding would help to account for the falling temperature and might also produce conditions suitable for the invasion of the area by the plexus of granite veins.

Later movements

A number of small late faults has been recognised in mapping the Gneissic Group. These are about vertical and trend approximately N.E.—S.W. The biggest fault which runs from the S.W. corner of Sokumvatn to Langvatn has an apparent downthrow of about twenty metres to the S.E. The other faults may have apparent throws either to the S.E. or N.W. but the sense of movement becomes more consistent if they are regarded as wrench, rather than as normal, faults. Almost all the faults then appear to be dextral (fig. 12).

The interpretation as wrench faults is strengthened by the fact that prominent vertical shear joints with the same trend are developed in numerous late granite, pegmatite and quartz veins (fig. 12).

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[1] Sillimanite (fibrolite) of an earlier generation is also present in these rocks.

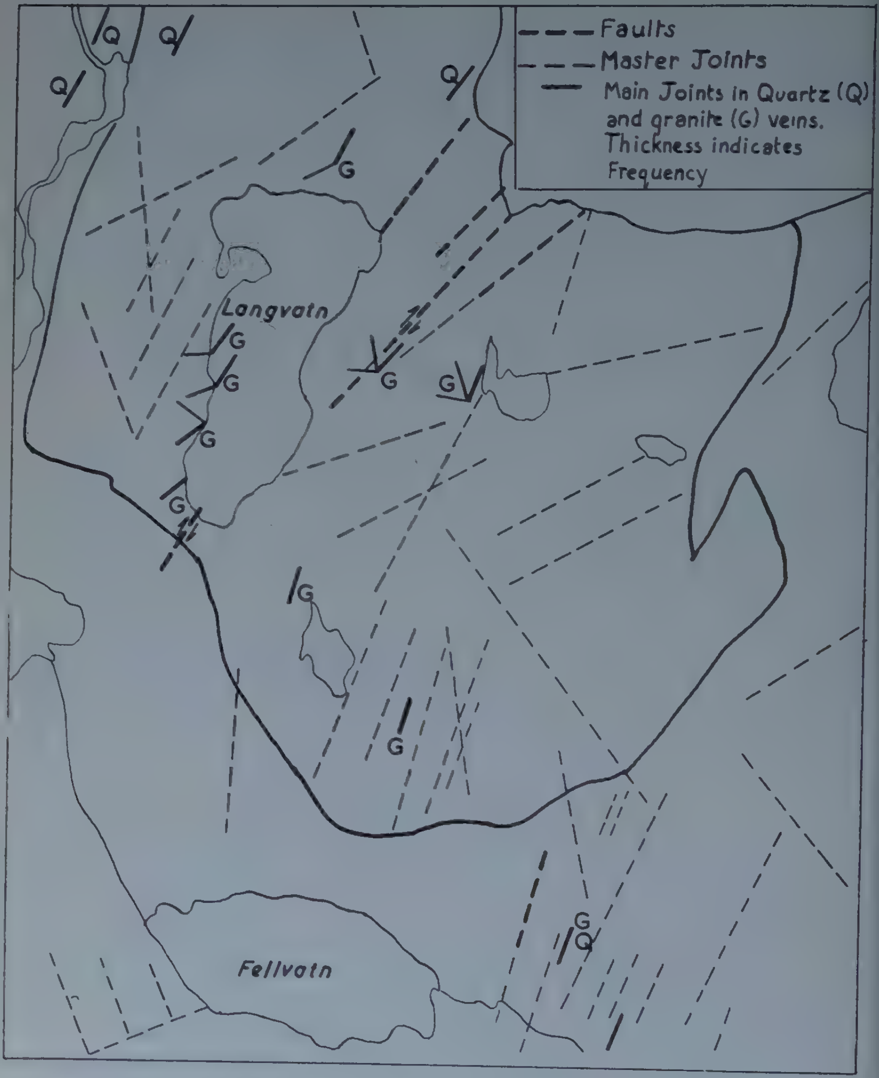
Faults
Master Joints
Main Joints in Quartz (Q) and granite (G) veins. Thickness indicates Frequency
Q
G
Langvatn
Fellvatn

Fig. 12. Relations of master joints, shear joints and wrench faults.

In some of the pegmatites west of Fellvatn the muscovite is orientated in bands parallel to these shear joints, almost at right angles to the trend of the dykes. Thus it seems that the last phase of movement represented by the faults and shear joints was contemporaneous with some of the last stages of crystallisation in the pegmatites.

The faults are also parallel to one of the most important master joint directions in the area, and many of the master joints must be regarded as incipient shear surfaces. In some vertical exposures single steep joint partings in the marbles have been seen to pass into granite boudins as sets of closely spaced shear joints with slightly different attitude to the parent joint. Other master joints, however, are related to the fold geometry. South-west of Fellvatn, the two sets of master joints at right angles may be regarded as 'ac' and 'bc'. The 'bc' joints are not strictly parallel to the measured first fold lineations and they may be related to the suggested second folding on N.E. axes rather than to the first folding. North of Sokumfjell prominent master joints are clearly 'ac' to the second folding. The joints have not been studied in detail in the field, but many other sets occur as well as those parallel to the master joints recognised on aerial photographs. They can often be tentatively correlated with the movement picture of one or other of the episodes of deformation, but their age is generally in doubt.

Concluding discussion

The main conclusions of the preceding sections are briefly reiterated here:

(1) There have been two main episodes of folding and a later minor episode of faulting in the Sokumvatn Area.

(a) an earlier more plastic deformation producing recumbent isoclinal folds on E.N.E.—W.S.W. axes. The Staburfjell Slide Zone allowed the Marble and Gneissic Groups to behave independently.

(b) a later less plastic deformation on approximately N—S axes, producing the Sokumvatn Synform which involves both the Marble and the Gneissic Groups. The present sheet dips are largely dependent on this episode.

(c) a non-plastic deformation producing N.E.—S.W. trending wrench faults.

(2) All the major overfolds, viz. the Bjørnhatten, Harefjell, Sokumfjell and Fiskevatn Folds were produced by the first folding. Their present plunges result from rotation in the Sokumvatn Synform during the second folding.

(3) Minor structures have been produced in both fold episodes.

(a) in the gentle western limb of the Sokumvatn Synform most of the minor structures belong to the first period of folding and are parallel to the major fold axes.

(b) In the steep eastern limb of the Sokumvatn Synform most of the minor structures belong to the second folding and are not related to the major overfolds. They vary greatly in direction because of pre-existing variations in the structures on which they were super-imposed.

(4) The episodes of deformation are distinctly separate in time. The first folding took place at the height of the regional metamorphism and before most of the granite emplacement. The second folding took place at a lower temperature after the emplacement of the soda granites; and the non-plastic faulting probably occurred during the last stages of consolidation of potash pegmatites and quartz veins.

A map illustrating the main features of the structural interpretation is given in fig. 11, and sketch sections across the area, based on this interpretation, are given in fig. 2. The E—W sections (A—D) are drawn roughly parallel to the axes of the first folding. They do not therefore, cut the first fold closures and they indicate the form of the Sokumvatn Synform produced in the second folding. The other section is drawn normal to the axes of the first folding and roughly parallel to the axes of the second folding. It therefore shows clearly the first fold structures.

It is not yet possible to correlate to the rock units and structural history here described with those recognised by KAUTSKY (1953) for the Sulitelma region, but some general comments may be made. KAUTSKY (1946) clearly regards the Sokumvatn Area along with most of the surrounding areas as part of the Large Seve Nappe which he suggests has been thrust hundreds of kilometres from the west over the Swedish Foreland. He recognises that in Norway the thrusting has been followed by a folding, producing domes and depressions, associated with remobilisation of the Basement granites. This folding can be reasonably correlated with the second folding of the Sokumvatn

Area, which has produced the Sokumvatn Synform and which is associated with granitisation. KAUTSKY (1947, p. 108) also believes, however, that it is this granitisation which has made the relatively plastic folding possible, and that, before the folding, the structural pattern was the same on both sides of the Norwegian frontier. This view is difficult to accept. The main thrusting of the Large Seve Nappe in the Sulitelma area was a non-plastic post-metamorphic deformation involving mylonitisation. In the Sokumvatn Area, on the other hand the second folding was preceded by an even more plastic syn-metamorphic deformation with the production of isoclinal overfolds. This first folding could be correlated with the post-metamorphic thrusting in the Sulitelma area by making the reasonable assumption that a single phase of deformation was affecting different metamorphic environments. But this is probably too simple a view. KAUTSKY (1953) has shown that the Large Seve Nappe is made up of a number of smaller units which had a complex metamorphic and structural history before the Large Seve Nappe as a whole was thrust over the Akkajaure Complex. Much more work is therefore needed before any definite conclusions can be drawn. In the meantime the writer is reluctant to accept the view that the Sokumvatn area is part of a great far travelled nappe.

Comment may also be made on the absence, in the Sokumvatn Area, of pronounced linear structures parallel to the local movement direction. Linear structures normal to the geosynclinal margin and regarded as 'a' lineations appear to occur consistently along the whole length of the Caledonian geosyncline in Norway. Often the evidence appears quite conclusive (KVALE 1948, OFTEDAHL 1948). It seems likely that such 'a' linear structures are characteristic of the thrust zones near the margins of the geosyncline where the basement is not far below. The thrusts, with which the 'a' structures are associated, are usually planar, post-metamorphic dislocations. The Sokumvatn Area, on the other hand, is representative of a different structural environment in the central zones of the geosyncline. The slides are syn-metamorphic, and the minor structures associated with them are b-structures. In the writer's opinion, therefore, there is not necessarily any conflict between the interpretation advanced here for the Sokumvatn Area and those advanced for other areas in Norway.

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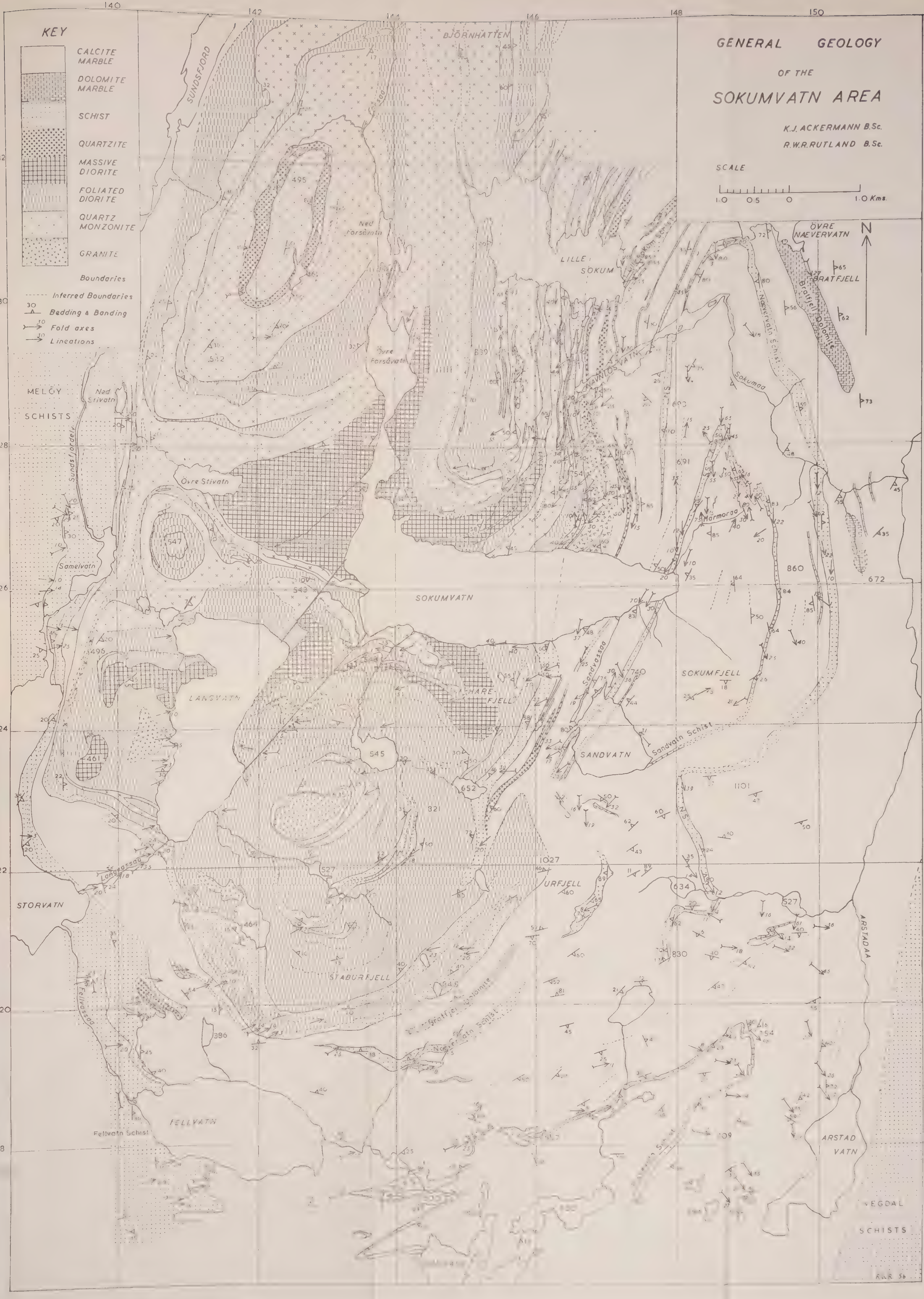

Plate 1. General Geological Map of the Sokumvatn Area.

CONTRIBUTIONS TO THE MINERALOGY OF NORWAY

No. 3. Bavenite from Boksjøen mineral mine, near Aspedammen in the county of Østfold.

By

Henrich Neumann and Thor L. Sverdrup.[1]

In 1952 Dr. O. J. Adamson visited the Boksjøen mineral mine in search of beryl and found on that occassion rosettes of a fibrous mineral which he could not find time to examine himself, and therefore passed over to the Geological Museum for identification. Dr. Adamson and the senior author revisited the place in June 1953 to collect additional material for further investigations of the then unknown mineral, which was later identified as bavenite.

The Boksjøen mineral mine (Boksjøen mineralgrube), which actually is a small quarry, is situated about 1 km SSE of the farm Toklund and about 200 m from the shore of the lake Nordre Boksjø in the parish of Idd in the county of Østfold, see key map fig. 1. In the pegmatite dyke which is 5—10 m wide striking between NE and NNE and dipping about 35° E, the following minerals have been found: quartz, alkali feldspar, plagioclase, mica, garnet, beryl, and bavenite. The dyke has been worked for feldspar, yielding some beryl as a by product.

The bavenite is found as radiating flat sheafs of lathshaped, grey crystals on the surface of or in cracks in beryl. The irregular sheafs may attain a size of 2—3 cm in diameter but are usually somewhat smaller. Muscovite (structure variety 2 M_1) is the only mineral found together with and occurring in the same way as bavenite, and is obviously paragenetical with it. Muscovite is older than bavenite.

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[1] Now at Norges Geologiske Undersøkelse, Oslo.

The optical properties are: 2 V + = 47° ± 3°, $\alpha = 1.580$, $\beta = 1.584$, $\gamma = 1.590$ (Average values of data published earlier for bavenites from other localities (6) : $\alpha = 1.581$ (1.578 — 1.586), $\beta = 1.583$ (1.579 — 1.588), $\gamma = 1.589$ (1.583 — 1.593)). An X-ray powder pattern taken with Mn-filtered Fe-radiation on a 9 cm camera is in every detail identical to the patterns of bavenite from Baveno, Italy and of bavenite (duplexite) from Londonderry, Australia. A picture of the bavenite pattern and X-ray powder data are published earlier by M. FLEISCHER and G. SWITZER (6).

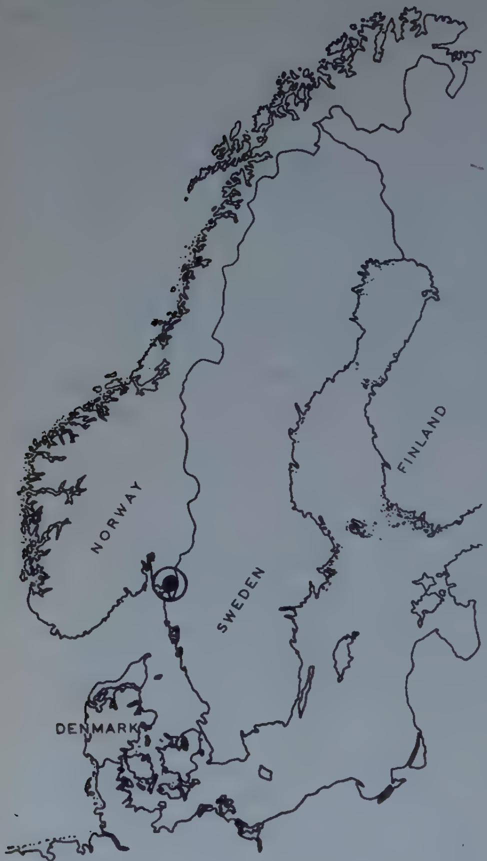

Fig. 1. Key map. The dot with a ring around it indicates the location of Boksjøen mineral mine.

Finds of bavenite have been recorded earlier:

1) From the Baveno granite, Italy, on feldspar crystals in pegmatitic druses together with younger mica, epidote, and laumontite. (1).

2) From Himalaya mine, Mesa Grande, San Diego Co., California in pegmatite as a pseudomorph after beryl, and also as fibrous spherules in cavities in relics of the original beryl crystal. (2).

3) From below the Muotta Nera, Piz del Laiblau, Val Nalps, Tavetschthal, Graubünden, and on the slopes of the east side of Val Casaccia, Val Cristallina, Graubünden in Switzerland, from alpine veins as radiating aggregates of platy fibres. (3).

4) From the Izumrudnaja emerald mines in the Malischevski district in Russia, on the dumps from the Kirov shaft, in the Skutinai shaft, and in shaft no. 5, partly together with microcline (with a con-

tent of 1.2 % BeO) replacing beryl or even as pseudomorphs after beryl, partly in small crosscutting veins either in the paragenesis: plagioclase, fluorite, apatite, corundophilite, muscovite, ripidolite, and alkali-feldspar (with .58 — 1.2 % BeO) or in the paragenesis: plagioclase, epidote, sphene, and prochlorite. Bavenite is a very young mineral in the parageneses, the only younger mineral is ripidolite which is found as small spherulites on the surface of crystal aggregates of bavenite. In the zone of weathering bavenite is unstable and is in turn replaced by bertrandite and some other unidentified minerals.

E. KUTUKOVA (4) has described this occurrence of bavenite and discussed its genesis at some length. He reached the conclusion that it was formed by reaction of plagioclase and beryl with late hydrothermal solutions. A minor part of the beryllium going into solution by this reaction was fixed again in situ by the formation of bavenite and microcline with a content of 1.2 % BeO, while the major part of it was removed and redeposited elsewhere as bavenite in small crosscutting veins together with the minerals mentioned above.

5) From Londonderry feldspar quarry near Coolgardie, Western Australia, in a pegmatite. Bavenite is "one of the last (minerals) to crystallize from the mineralising solution penetrating the pegmatite and ..... has crystallized at the same time as or later than the idiomorphic quartz crystals". It is older, though, than bityite which occurs together with it and is probably genetically closely related to it. (5).

6) From Rutherford mine, Amelia, Virginia, in a pegmatite as rosettes of thin platy crystals from a cavity in cleavelandite. A few minute crystals of younger bertrandite are emplanted on the bavenite. (6).

In the Boksjøen occurrence bavenite is found as the very youngest mineral. It is highly improbable that the fairly large crystals of bavenite and of muscovite were formed by weathering even if this cannot be disproved beyond doubt with the material at the authors' disposal and with the present shallow exposure of the pegmatite dyke. In harmony with data cited above from the other known bavenite occurrences the authors feel confident in stating that the Boksjøen bavenite is a deuteric mineral formed by the reaction of a late pegmatitic hydrothermal solution with beryl which apparently during that late stage of the pegmatite formation became unstable under the prevailing

PTX conditions. The necessary amounts of Ca and K to form bavenite and muscovite seem to have been supplied by the hydrothermal solution at play as we have no indication of a contemporaneous reaction of plagioclase to supply Ca as suggested by Kutukova in the case of the Izumrudnaja bavenite. The balance of Al and Si is hard to calculate without further data on the relative amounts of paragenetical bavenite and muscovite and without an exact knowledge of the formula of bavenite.

Acknowledgement

Thanks are due to Mrs. Natascha Heintz for translating Kutukovas paper (4).

Added in proof: The mineral pilinite from Striegau, Silesia (A. v. Lasaulx. Jb. Min. 1876, p. 358) has recently been found to be identical with bavenite. (Brian Mason, 1959, personal communication).

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Geologisk Museum Oslo University August 21, 1959.

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DISTRIBUTION OF Ba AND Sr IN MICROCLINE IN SECTIONS ACROSS A GRANITE PEGMATITE BAND IN GNEISS

BY

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Abstract. Spectrochemical determinations of Ba and Sr in microcline samples taken along 5 sections across a granite pegmatite band in plagioclase gneiss exhibit highly variable concentrations. Particularly low concentrations appear to occur within definite parts, which have apparently formed at later stages than the rest of the pegmatite.

A contour map showing roughly the Ba distribution within the examined part of the pegmatite band has been tentatively constructed.

This is a supplement to an earlier paper by the writer (1). That paper reported on spectrochemical determinations of Ba and Sr in feldspars, and conclusions drawn from this data. Special attention was paid to the granite pegmatite veins and bands within a small area of the plagioclase gneiss complex in the island of Justøy on the Skagerrack coast. It was found that within a single larger pegmatite band, 2 to 3 meters wide, the Ba contents of the microcline may vary by a factor of more than 50, and the Sr contents somewhat less. It might be of interest to examine this particular pegmatite band somewhat more in detail. Last summer I collected a considerable number of microcline samples along 5 sections across the band, all within a length of about 30 meters. Spectrograms of these samples were prepared by Mr. F. Chr. Wolff in the laboratory of the Institute, the conditions — spectrographic equipment, size and shape of the carbon cathodes, current, exposure time, type of plates, development — being exactly the same as in the earlier investigations. The standard

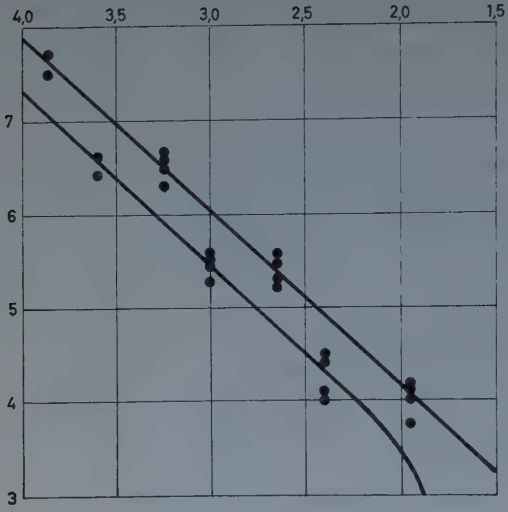

Fig. 1. Working curves. Upper: Ba 4554. Lower: Sr 4607. Dots represent intensity readings in the standard mixture spectrograms. Horizontal scale: concentrations of BaO and SrO in log. p.p.m. Vertical scale: no. of last step easily visible (due to rotating step sector).

mixtures, however, were new and improved. They were made with a Ba- and Sr-rich microcline which had been carefully analyzed for Ba and Sr by the chemist of the Institute, Mr. O. Hagberg, the base substances for 2 series of mixtures being as earlier amazonite and cleavelandite from Tørdal. The mixtures, including the pure analyzed microcline, contained the following concentrations: BaO — 0.72, 0.18, 0.045, 0.009 %; SrO — 0.40, 0.10, 0.025, 0.005 % (apart from a little Sr present in the amazonite). From duplicate spectrograms of each series the working curves shown in Fig. 1 were obtained. The plots of the individual intensity readings show that the curves may be used with some confidence. The method does not give a very high degree of accuracy: the individual determinations may possibly be in error by as much as ± 50 %; in general they are probably better than this. At any rate this is more than sufficiently accurate for the present purpose.

Comparison with the curves used in earlier investigations (2) shows that the difficulties then met with have now been largely removed. The earlier standard mixtures contained the compounds BaO and SrO, and it is now evident that these mixtures do not evaporate in the carbon arc in the same manner as feldspars containing Ba and Sr in their lattices do. The old working curves for Ba and Sr in amazonite intersected, and the Sr curves obtained from amazonite mixtures did not agree with those obtained from cleavelandite mixtures. The new curves for Ba and Sr are largely parallel, and they are not perceptibly influenced by the base substance (amazonite or cleave-

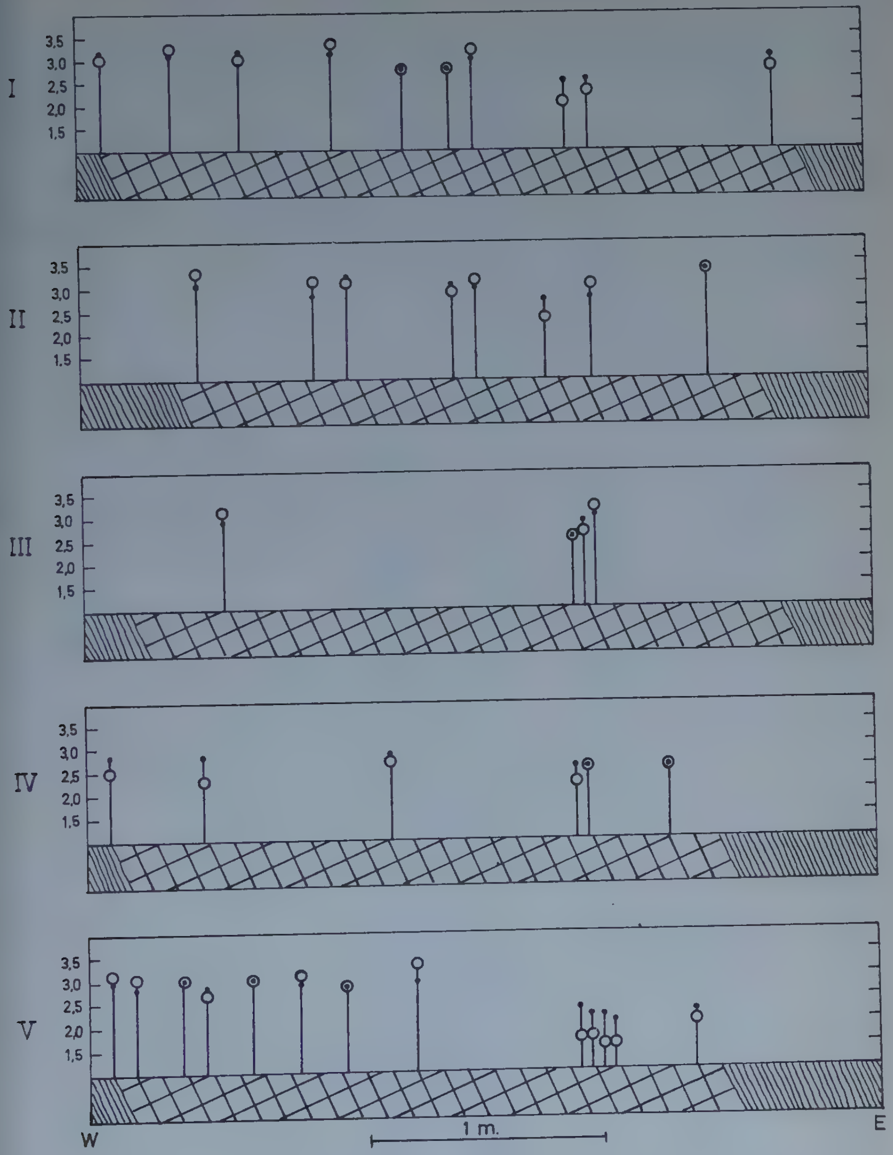

Fig. 2. Distribution of BaO (circles) and SrO (dots) in microcline in 5 sections across the pegmatite band. The gneiss is indicated left and right. Vertical scales: concentrations in log p.p.m.

landite) of the mixtures. Part of the discussion on this matter in the earlier paper (2) is therefore irrelevant. One conclusion to be drawn now is that the Sr line 4607 is more persistent when originating from an amazonite mixture containing the compound SrO than when originating from microcline containing Sr. In spite of this discrepancy the new curves do not give values notably different from the old ones, except in the case of Sr at low concentrations in microcline.

The results of the measurements are shown graphically in Fig. 2. In each of the 5 sections across the pegmatite band a number of ordinates show by their lengths the log p.p.m. of the BaO and SrO contents in microcline at the corresponding abscissae. It is seen that in each section there is a more or less marked drop in the contents near the eastern boundary of the pegmatite band. In section IV there are also low contents near the western boundary. Normally the BaO and SrO contents of the pegmatite microcline appear to oscillate somewhat about 1000 p.p.m., but in certain parts — containing usually particularly coarse and light coloured microcline — the contents may drop below 100 p.p.m. or even 50 p.p.m. for BaO and towards 100 p.p.m. for SrO. It is obvious that these latter microclines have not attained equilibrium with neighbouring microclines richer in Ba and Sr. They must have formed at late stages when the rate of diffusion of Ba and Sr was lower than the rate of crystallization. The interpretation given earlier (1) as to the origin and development of the pegmatites in this area I see no reason to modify in view of the new data.

Fig. 3 is a tentative contour map of the distribution of BaO in microcline in the examined part of the pegmatite band. (SrO is distributed in a similar way, but here the concentration variations are much less pronounced.) Only along the sections I to V the map is based on actual measurements. The contour lines connecting the sections are thus rather uncertain. They may in reality be much more irregular. This is evident from the fact that in some cases the BaO-content has been shown to vary greatly within a distance of a few centimeters; even within a single microcline crystal notable variations appear to be possible. Nevertheless the map gives a picture which must be correct in essentials. The "younger" Ba-poor microcline is found mainly within a zone which widens and narrows irregularly along the pegmatite band. Thus the shape of the Ba-poor part of the pegmatite is not that of a regular band, but it is still notably

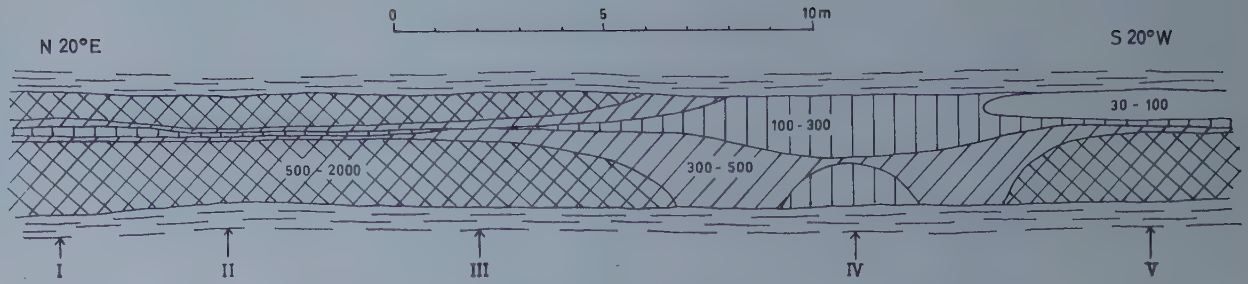

Fig. 3. Tentative contour map of the distribution of BaO in part of the pegmatite band. Position of sections I to V shown. The figures denote parts per million in microcline.

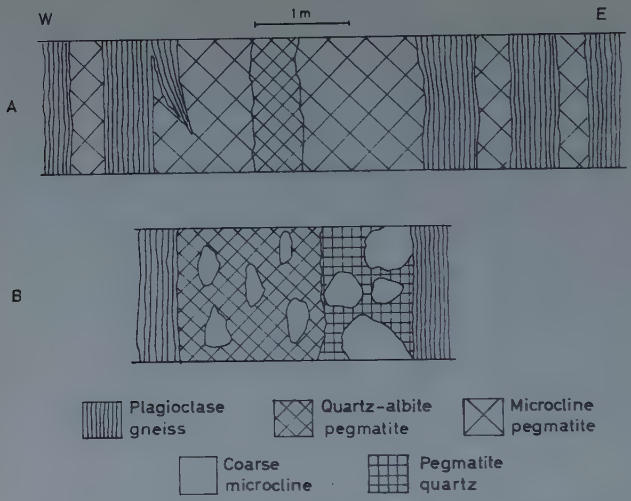

Fig. 4. Two somewhat schematic maps across the pegmatite band. A, near section II. B, at section V. Map A shows smaller companion pegmatite bands; the microcline of these belongs to the Ba- and Sr-rich type.

influenced by the tectonics of the area. The clusters of very coarse-grained and extremely Ba-poor microcline seem to have developed in complete absence of stress action; they are associated with massive pegmatite quartz. Much of the relatively Ba- (and Sr-) rich microcline occurs in the shape of patches within medium-grained quartz-plagioclase pegmatite, which in places forms apparently older parts of the pegmatite band (Fig. 4). The plagioclase (albite-oligoclase) of this kind of pegmatite tends to be somewhat poorer in Sr and Ba than the plagioclase of the country rock (plagioclase-quartz gneiss), but the differences are in general not very marked.

As pointed out above, equilibrium distribution of Ba and Sr between neighbouring feldspar crystals in this pegmatite is not a general rule. However, in some cases of microcline and plagioclase in close association there is a regularity which may possibly indicate equilibrium distribution:

		BaO, p.p.m.	SrO, p.p.m.
1.	Microcline	1500	1200
	Plagioclase	500	4000
2.	Microcline	1000	1200
	Plagioclase	600	2000

		BaO, p.p.m.	SrO, p.p.m.
3.	Microcline	1000	1000
	Plagioclase	500	2000
4.	Microcline	5000	500
	Plagioclase	500	5000

Number 4 represents 3 pairs taken from contacts between plagioclase of the gneiss and microcline of other, much smaller, pegmatite bands in the same area. In all of these examples the coefficient of distribution of BaO between microcline and plagioclase is seen to be very nearly equal to that of SrO between plagioclase and microcline, even if the actual concentrations vary considerably. In one or two other examined cases the distributions do not follow this rule; this may indicate that in these cases equilibrium has not been reached.

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NOTISER

Montmorillonitt fra Skyrvedalen i Hemsedal.

Montmorillonite from Skyrvedalen in Hemsedal, S. Norway.

Av
I. TH. ROSENQVIST

Abstract: A highly plastic montmorillonite breccia rock from Skyrvedalen, Hemsedal, Central Norway, has been investigated by means of X-ray and chemical methods. The rock is found to consist of a nearly hundred percent calcium dioctahederic montmorillonite of extraordinary purity.

Under kartleggingsarbeidet i forbindelse med utbyggingen av Oslo Lysverkers kraftanlegg i Hemsil I, ble det påtruffet en meget ren montmorillonittbreksje i driftstunnelen under Skyrvedalsskaret i Hemsedal. Forekomsten av denne montmorillonitt minner meget sterkt om den montmorillonitt man fant i Fortun i Sogn, omtalt i Norsk Geologisk Tidsskrift, bind 37, s. 403—414. Den skiller seg imidlertid i visse henseender fra dette mineral, i det vesentlige i sine røntgendata, men også i sin kjemiske sammensetning. Mineralet fra Skyrvedalen i Hemsedal var ualminnelig rent, om mulig ennå renere enn montmorillonitten fra Fortun. Mikroskopisk undersøkelse av utplukkede transparente bergartsstykker viste etter knusing *ingen* andre mineraler enn leirmineralene. Den kjemiske analyse utført ved Sentralinstituttet for Industriell Forskning ga følgende resultat, sammenlignet med montmorillonitten fra Fortun, begge omregnet til preparat tørket ved 110° C.

Tabell 1

	Skyrvedalen	Fortun
Glødetap		
110–950° C	9,61 %	11,54 %
SiO_2	59,58 %	57,50 %
Al_2O_3	22,96 %	21,60 %
Fe_2O_3	0,47 %	0,94 %
MgO	3,67 %	4,20 %
CaO	3,38 %	3,92 %
Na_2O	0,06 %	0,11 %
K_2O	—	0,03 %
	99,73 %	99,95 %

Beregner man formelen på basis av denne analyse, får man følgende resultat:

Skyrvedalen

Oktahederiske lag.
(Octahederal layers)

$$\left.\begin{array}{ll}\text{Mg} & 0{,}36\\ \text{Al} & 1{,}62\\ \text{Fe}^{\text{III}} & 0{,}02\end{array}\right\}\begin{array}{l}2{,}00-0{,}36 \text{ valenser}\\ (\text{OH})_2\end{array}$$

Tetrahederiske lag
(Tetrahederal layers).

$$\left.\begin{array}{ll}\text{Si} & 3{,}87\\ \text{Al} & 0{,}13\end{array}\right\}\begin{array}{l}4{,}00-0{,}13 \text{ valenser}\\ \text{O}_{10}\end{array}$$

$-0{,}49$ valenser

Uttbyttbart.
(Exchangable)

$$\left.\begin{array}{ll}\text{Ca} & 0{,}24\\ \text{Na} & 0{,}01\end{array}\right\}\text{Tilsv. } 0{,}49 \text{ valenser}$$

1,078 H_2O (110° C)

Fortun

Oktahederiske lag
(Octahederal layers)

$$\left.\begin{array}{ll}\text{Mg} & 0{,}42\\ \text{Al} & 1{,}55\\ \text{Fe}^{\text{III}} & 0{,}04\end{array}\right\}\begin{array}{l}2{,}01-0{,}39 \text{ valenser}\\ (\text{OH})_2\end{array}$$

Tetrahederiske lag.
(Tetrahederal layers).

$$\left.\begin{array}{ll}\text{Si} & 3{,}84\\ \text{Al} & 0{,}16\end{array}\right\}\begin{array}{l}4{,}00-0{,}16 \text{ valenser}\\ \text{O}_{10}\end{array}$$

$-0{,}55$ valenser

Utbyttebart.
(Exchangeable

$$\left.\begin{array}{ll}\text{Ca} & 0{,}26\\ \text{Na} & 0{,}01\end{array}\right\}\text{Tilsv. } 0{,}53 \text{ valenser}$$

1,6 H_2O (110° C)

Begge preparater svarer til utpregete dioktahederiske montmorillonitt beidelitt-mineraler.

Røntgenundersøkelse ble foretatt med spectrometer og Guinierkamera, dels på det naturlige mineral, dels på glyserolbehandlet preparat og til sist på natriumpyrofosfatbehandlet preparat. Disse undersøkelser ga følgende resultat: (*tabell 2*).

Tabell 2

Ind.	$Na_4P_2O_7$-beh. I W 700%. (Guinier) d(Å)	I	$Na_4P_2O_7$-beh. II W 700%. (Guinier) d(Å)	I	Naturl. mineral (Spectrometer) d(Å)	I	Glyserolbehandl. (Guinier) d(Å)	I	Glyserolbehandl. (Spectrometer) d(Å)	I
001	Ca 15,0	S			14,97	VVS	18	VVS	18,0	VVS
002	Na 18,2	VS	Na 19,19	VS			8,7 ?	S	9,01	M
hkl									8,04	VW
hkl									7,08	VW
003	Na 12,3	S			5,01	S	5,8	M	5,90	M
hkl									4,40	VS
hkl									3,98	W
hkl									3,83	W
110(asm)					4,50	VS				
004	Na 9,2	VW	Na 9,56	S	3,77	W				
hk0					3,50	VW				
hk0					3,30	VW				
hkl(004 Ca?)	3,63	VW	6,70	VW						
005	Ca 2,96	W			3,02	S	3,5	M	3,54	M
006	Na 6,18	M	Na 6,27	M	2,58-2,50	M	2,92	W	2,95	W-M
hkl	5,9	W			2,26	VW				
007	Ca 2,12	VW			2,15	VW	2,55	S	2,55	M
hkl							2,47	M		
hkl							2,36	W		
hkl									2,49	
hkl									2,46	VW
hkl									2,44	
hkl									2,37	
200	Na 4,47	VS	Na 4,50	S	1,88	VW			2,25	W
008			Ca 1,83	VW						
hkl			3,69	W	1,70	V-M	2,10	VW	2,14	
									2,10	VW

hk0			1,69	S			1,69	M	1,70	W-M
hkl							1,62	VW		
009					1,50	M-S				
0012	Na 3,07	M	3,11 ?	S	1,493	M-S	1,493	M-S	1,50	M
hkl					1,285	W				
0013	Na 2,76	M	2,79	VW						
hkl	2,56	S	2,58 2,53	M						
hkl	2,35	VW	2,31	W						
hkl			1,70	VW						
0013					1,243	W				
0019	Na 1,95	VW								
(0024Na 0010Ca)	Na 1,49	VS	1,50	M						

I = VVS Ualminnelig sterk.
I = VS Meget sterk.
I = S Sterk.
I = M Middels.
I = W Svak.
I = VW Meget svak.

Ved glyserolbehandling ekspanderte mineralet fra en basisplanavstand på ca. 15 Å til ca. 18 Å. Dette er vanlig for kalsium-montmorilloniytt. Ved natrium-pyrofosfat-behandling, hvorved mineralet overføres fra kasiumtilstand til natriumtilstand, synes basisplanavstanden å være øket til 36—38 Å i fullstendig vannmettet prøve. Dette bygger på at d.verdiene 18,2 og 19,7 synes å representere refleks 002 ved flytegrensen, dvs. ca. 700 % vann i forhold til tørrvekt. Bergartens flytegrense i naturlig tilstand var 125,1 %, i natriumtilstand 722 %.

Forekomstens mektighet er ialt ca. 8 m i nær vertikalstående lag med tykkelser varierende mellom noen få centimeter og en halv meter. Bergarten viste utpreget svelning etter tunnelgjennomslaget, idet dens naturlige vanninnhold vare bare ca. 20 % og dens plastisitetsgrense 55. Svelningstrykket mot utstøpningen antas derfor å kunne bli betydelig (av størrelsesorden 10 atmosfærer og derover).

Manuskript mottatt 28. mai 1959.
Trykt desember 1959.

NY LITTERATUR

CONRAD BURRI: *Petrochemische Berechnungsmetoden auf äquivalenter Grundlage.* 337 sider, 66 figs., Sch. fr. 38, Birkhäuser 1959.

Alle som driver petrografisk forskning i dag må beregne sine bergartsanalyser. Disse beregninger bør utføres på grunnlag av kationprocenter, ikke som tidligere gjort, på grunnlag av vektsprocenter. Dette er det som menes når det i tittelen står «auf äquivalenter Grundlage».

Boken, som gir en meget grundig innføring i disse metoder består av 4 deler:

I. Bakgrunnsstoff for forståelse av bergartsanalysen, dens nøyaktighet og anvendelse. En oversikt over de forskjellige beregnings- og projeksjonsmetoder; særlig inngående blir grafiske fremstillinger behandlet.
II. Petrokjemiske beregninger etter P. Niggli; verdiene *si, al, fm, c, alk, k, mg, ti,* usw; ekvivalentnormene.
III. Forslag av Barth og Eskola om å ta anionene i betraktning; Barth's standardcelle og dens anvendelse for beregning av stoffbytte ved isovolumetriske metasomatiske processer.
IV. Tabeller over basis- og ekvivalentnormative forbindelser. Nyberegnede tabeller over ekvivalent-tall (kation-proporsjoner) av alle bergartsdannende oxyder hvorved samtlige utregninger lettes betydelig for alle som har boken.

Det blir lagt stor vekt på fullstendig gjennomregnede og gjennomdiskuterte eksempler av både eruptive og metamorfe bergarter.

T. F. W. Barth.

NORSK GEOLOGISK FORENING

Møte nr. 381, 26. februar 1959.

Institutt for Geologi, Blindern.

Til stede 22 medlemmer, 12 gjester.

Referat fra generalforsamlingen og møte nr. 380 ble godkjent.

Følgende innvalg ble godkjent:

436. Cand. mag. Jens Hysingfjord, Institutt for Geologi, Blindern. Foreslått av H. Skålvoll og K. S. Heier.

Foredrag av direktør *Kåre Landmark:* «Geologi i midt-Troms».

I ordskiftet etter foredraget deltok:

W. Elders, Chr. Oftedahl, O. Holtedahl, T. Barth, T. Strand, F. Vokes, P. Holmsen, S. Føyn, B. Dietrichson, G. Liestøl, E. Kiil og foredragsholderen.

Møte nr. 382, torsdag 9. april 1959.

Drammensveien 78, Oslo.

Til stede 32 medlemmer, 10 gjester.

Referat fra forrige møte ble godkjent. Formannen mintes bergingeniør Henning Marstranders bortgang.

Følgende innvalg ble godkjent:

437. Cand.mag. Olav H. J. Christie, Geologisk Museum, Oslo. Foreslått av H. Neumann og K. S. Heier.
438. Cand.mag. Sveinung Bergstøl, Geologisk Museum, Oslo. Foreslått av H. Neumann og K. S. Heier.
439. Professor R. V. Dietrich, Virginia Polytechnic Institute, Blacksburg, Va., U.S.A. Foreslått av T. Barth og H. Neumann.
440. Dr. Roger David Morton, Geologisk Museum, Oslo. Foreslått av H. Neumann og K. S. Heier.

441. Wilfred Allan Elders, B. Sc., 47, Viador, Chesterle-Street, County Durham, England. Foreslått av H. Neumann og K. S. Heier.
442. Stevan Divljan, Federal Geological Institute, Rovinjska 12, Belgrade, Yugoslavia. Foreslått av H. Neumann og K. S. Heier.

Foredrag av professor *Anders Kvale:*
«Strukturundersøkelser i Gotthardområdet.»

I ordskiftet etter foredraget deltok:
T. Barth, Chr. Oftedahl, T. Gjelsvik, T. Strand og foredragsholderen.

Mineralogisk Gruppe av Norsk Geologisk Forening.

På møte nr. 375 i Norsk Geologisk Forening ble det gjort følgende vedtak:

«Norsk Geologisk Forening samtykker i at ovennevnte forslagsstillere (Professor I. Oftedal, professor T. F. W. Barth og dr. H. Neumann) og andre interesserte medlemmer danner en mineralogisk gruppe innen foreningen». Det ble forutsatt at opprettelsen av gruppen ikke skal medføre økede utgifter for foreningen.

En komite bestående av: Professor I. Oftedal, professor T. F. W. Barth, dr. H. Neumann og cand.real. K. S. Heier innbød til konstituerende møte i Drammensveien 78, torsdag 8/10 1959 kl. 18.30.

Møtet ble ledet av cand.real. K. S. Heier, som sekretær i Norsk Geologisk Forening. Dr. H. Neumann orienterte om årsaken til gruppens dannelse. Den ovennevnte komite hadde utarbeidet følgende lovforslag som ble vedtatt:

Navn: *Mineralogisk Gruppe*
av Norsk Geologisk Forening.

§ 1. Den Mineralogiske Gruppe skal søke å fremme den mineralogiske forskning i Norge og delta i det internasjonale samarbeid bl. a. gjennom medlemskap i «The International Mineralogical Association».

§ 2. Medlemmene av gruppen skal være mineralogisk interesserte medlemmer av Norsk Geologisk Forening; nye medlemmer innvelges av gruppens styre.

§ 3. Styret består av formann, sekretær og en fast representant til The International Mineralogical Association.

§ 4. Styrets medlemmer velges ved skriftlig stemmegiving i oktober måned hvert tredje år.

§ 5. Medlemmer til kommisjoner i I.M.A. og til eventuelle andre lignende funksjoner oppnevnes av styret.

Til styre ble valgt:
Formann, professor I. Oftedal.
Sekretær, cand.real. K. S. Heier.
Fast representant til I.M.A., dr. H. Neumann.

Styret har oppnevnt følgende medlemmer som representanter i kommisjoner tilknyttet I.M.A.:

Commission on abstracts, professor I. Oftedal.
—»— mineral data, professor I. Oftedal.
—»— new minerals, dr. H. Neumann.
—»— museums, dr. H. Neumann.

Det har pr. 8/10—1959 meldt seg 28 medlemmer.

Dr. H. Neumann representerte gruppen på møtet av «The International Mineralogical Association» i Zürich i august/september d. å.

Mineralogisk Gruppe av Norsk Geologisk Forening.

Adamson, Olge J., dr. Königlich Norwegische Botschaft, Drachenfelsstrasse 9, Bonn.
Barth, Tom. F. W., prof., Geologisk Museum, Oslo N.Ø.
Bergstøl, Sveinung, cand.mag., Geologisk Museum, Oslo N.Ø.
Bjørlykke, Harald, dir., Norges Geologiske Undersøkelse, Josefinegt. 34, Oslo N.V.
Bruun, Brynjolf, ing., Geologisk Museum, Oslo N.Ø.
Bryn, Knut O., statsgeolog, Norges Geologiske Undersøkelse, Eilert Sundtsgt. 32, Oslo.
Bugge, Jens, A.W. prof., Geologisk Institutt, N.T.H., Trondheim.
Christie, Olav, H. J., cand.real., Geologisk Museum, Oslo N.Ø.
Dons, Johannes A., førstekonservator, Geologisk Museum, Oslo N.Ø.
Egeberg, Ferd. P., berging., Rådhusgt., 5 B, Oslo.
Heier, Knut S., cand.real., Geologisk Museum, Oslo N.Ø.
Hernes, Ivar, cand.real., Geologisk Institutt, Bergen.
Ihle, Ole K., cand.real., Norges Geologiske Undersøkelse, Josefinegt. 34, Oslo N.V.
Liestøl, Gerd B., cand.real., Institutt for Geologi, Blindern.
Låg, J., prof., Norges Landbrukshøgskole, Vollebekk.
Mortenson, Magne, prof., Norges Tekniske Høyskole, Trondheim.
Neumann, Henrich, dr., Geologisk Museum, Oslo N.Ø.
Nilssen, Borghild, cand.mag., Geologisk Museum, Oslo N.Ø.
Oftedahl, Christoffer, prof., Geologisk Institutt, N.T.H., Trondheim.
Oftedal, Ivar, prof., Institutt for Geologi, Blindern.
Reitan, Paul H., statsgeolog, Norges Geologiske Undersøkelse, Geologisk Museum, Oslo N.Ø.
Rosenqvist, Ivan Th., prof., Norges Geotekniske Institutt, Forskningsveien 1, Blindern.

Skålvoll, Harald, cand.mag., Norges Geologiske Undersøkelse, Josefinegt. 34, Oslo N.V.

Strand, Trygve, prof., Institutt for Geologi, Blindern.

Sverdrup, Thor L., statsgeolog, Norges Geologiske Undersøkelse, Josefinegt. 34, Oslo N.V.

Sæbø, Per Chr., cand.mag., Norges Geologiske Undersøkelse, Josefinegt. 34, Oslo N.V.

Thorkildsen, Chr. D., cand.real., Norges Geologiske Undersøkelse, Geologisk Museum, Oslo N.Ø.

Vokes, Frank M., dr., Geologisk Museum, Oslo N.Ø.

Geologisk ekskursjon til Tangenåsen, Sørkedalen.

Norsk Geologisk Forening arrangerte ekskursjon til Tangenåsen, Sørkedalen, torsdag 7. mai 1959.

Avreise med buss fra Rådhusplassen kl. 10.05 til Zinoberkrysset. Derfra tur til Tangen gård, over Tangenåsen og ned til Sørkedalen. Ekskursjonen ble avsluttet ca. kl. 16.00.

Ekskursjonens formål var å demonstrere felsitporfyrer og deres dannelse ble diskutert.

Ledere var: Professor O. Holtedahl og professor Chr. Oftedahl.

Med på ekskursjonen var 21 medlemmer og 14 gjester.

GEOLOGNYTT

Norges geologiske undersøkelse

Dr. philos. Tore Gjelsvik fratrådte som statsgeolog I 1. juli 1959.

Dr. philos. Christoffer Oftedahl fratrådte som statsgeolog I 1. juli 1959.

Statsgeolog II Rolf W. Feyling-Hanssen ble beskikket som statsgeolog I fra 1. juli 1959.

Cand. real. Johannes Færden ble beskikket som statsgeolog I fra 1. oktober 1959.

Vitenskapelig assistent Knut Ørn Bryn tiltrådte som statsgeolog II 1. juli 1959.

Cand. real. Christian Dick Thorkildsen tiltrådte som midlertidig vitenskapelig assistent 15. mars 1959.

Cand. mag. Per Christian Sæbo tiltrådte som midlertidig vitenskapelig assistent 15. april 1959.

Midlertidig vitenskapelig assistent Ole-Kristoffer Ihle ble fast ansatt i stillingen fra 1. juli 1959.

Statsgeolog I Per Holmsen tiltrådte sin stilling som statsgeolog I etter endt permisjon 1. april 1959.

Universitetet i Oslo

Institutt for geologi

Professor Ivar Oftedal har 1. januar 1959 avløst professor Leif Størmer som instituttets bestyrer.

Professor Leif Størmer har frisemester høsten 1959, og benytter dette til en studietur i Syd-Europa.

Størmer er utnevnt til representant for Universitetet i Oslo ved den 21. internasjonale geologkongress i Norden 1960.

Vitenskapelig assistent Jens Hysingjord har permisjon i et år fra 1. august 1959.

Cand. real. Fredrik Chr. Wolff er ansatt som vikar for Hysingjord.

Som midlertidig preparant er ansatt Rolf Larsen.

Mineralogisk-geologisk museum

Professor Tom. F. W. Barth er valgt som president for International Geochemical Society.

Førstekonservator dr. HENRICH NEUMANN er utnevnt som bestyrer av museet fra 5. desember 1958.

Konservator JOHANNES A. DONS ble 1. juli 1959 ansatt som førstekonservator samme sted.

Cand. mag. SVEINUNG BERGSTØL er ansatt som midlertidig vitenskapelig assistent ved museet fra 1. juli 1959.

Frk. JORUNN WIIK er ansatt som teknisk assistent ved røntgenlaboratoriet fra 1. august 1959.

Følgende gjestende forskere arbeider ved museet året 1959/60:

Professor HAROLD D. WRIGHT, The Pennsylvania State University. U.S.A., Fulbrightstipendiat.

SCOTT B. SMITHSON, M.A. Wyoming University, U.S.A.

BRIAN PLAYLE, University of Nottingham, England, engelsk stipendiat.

Paleontologisk museum

Professor A. HEINTZ har som medlem av en norsk delegasjon besøkt paleontologiske og geologiske institusjoner i Moskva og Leningrad i oktober 1959. Innbydelsen var fra Vitenskapsakademiet i Sovjetsamveldet til Det Norske Vitenskapsakademi i Oslo.

HEINTZ var også invitert til Newcastle i september 1959 til åpningen av Department of Geology ved King's College, University of Durham, og holdt her et foredrag om devon på Svalbard.

Cand. real. SVEIN B. MANUM er ansatt som museumsstipendiat for året 1959/60, med stud. real. FRANK NIKOLAISEN som vikar en del av tiden.

Norsk Polarinstitutt

Cand. real. HARALD MAJOR har fra 1. januar 1958 bestyrt Industridepartementets undersøkelser av Statens kullfelter på Svalbard.

Cand. real. AUDUN HJELLE har fra 1. januar 1959 arbeidet midlertidig som geolog ved Norsk Polarinstitutt med arbeidsfelt: Fortsatt geologisk kartlegging av området mellom Isfjorden og feltet omkring Bellsund.

Cand. real. THOR SIGGERUD er midlertidig ansatt som geolog II ved Norsk Polarinstitutt fra 1. januar 1960.

Norges Statsbaner

Cand. real. FREDRIK CHR. A. HUSEBY tiltrådte 11. mai 1959 stillingen som jernbanegeolog ved Geoteknisk kontor, Hovedstyret for statsbanene, Oslo.

Universitetet i Bergen

Geologisk institutt

Professor ANDERS KVALE har friår fra 1. juli 1959 til 30. juni 1960.

Cand. real. INGE BRYHNI er ansatt som vitenskapelig assistent under Kvales friår, lønnet av Norges almenvitenskapelige forskningsråd.

Norges tekniske høgskole

Geologisk institutt

Dr. philos. CHRISTOFFER OFTEDAHL ble i statsråd 12. juni 1959 utnevnt til professor i geologi etter TH. VOGT. Han tiltrådte stillingen 1. juli 1959.

Cand. real. FREDRIK CHR. A. HUSEBY fråtrådte 11. mai stillingen som vitenskapelig assistent.

Geofysisk Malmleting

Cand. real. FINN J. SKJERLIE er beskikket som geolog I fra 1. juli 1958.

Geolog HANS-PETER GEIS er ansatt som midlertidig geolog II fra 1. mars 1958.

Eksamener

Ved Universitetet i Oslo er embetseksamen med mineralogi-petrografi som hovedfag fullført i vårsemesteret 1959 av:

OLAV H. J. CHRISTIE. Hovedoppgave: «De fysikalske forhold under intrusjonen av Egersundsområdets doleritt-ganger belyst ved kjemisk og petrografisk analyse, samt ved eksperimenter.»

FREDRIK CHRISTIAN WOLFF. Hovedoppgave: «Geologiske undersøkelser innen kartbladet Verdal.»

I høstsemesteret 1959 av:

ARNE W. GRONHAUG. Hovedoppgave: «En geologisk undersøkelse i området Majavatn—Kvigtind på kartbladet Børgefjell.»

MAGNE GUSTAVSON. Hovedoppgave: «En undersøkelse av de høymetamorfe bergarter på den nordvestlige del av kartblad Børgefjell.»

Doktorgrad

Statsgeolog PAUL H. REITAN er av Det akademiske kollegium ved Universitetet i Oslo blitt kreert til *doctor philosophiae* på grunnlag av sin avhandling: «Pegmatite veins and the surrounding rocks. I—IV.»

MEDLEMSLISTE

à jour oktober 1959.
[1]: korresponderende medlem.
*: livsvarig medlem.
Tallet i parentes er innvalgsåret.
(S), stifter (18. februar 1905).

*Adamson, Olge J., dr. Königlich Norwegische Botschaft, Drachenfelstrasse 9, Bonn, Tyskland. (1945).
Ahlmann, Hans W:son, professor, Fregattvägen 6, Gräsdal, Stockholm S.V., Sverige. (1919).
Andersen, Bjørn, univ.stip. Geologisk institutt, Blindern, Oslo. (1952).
Asklund, Bror, professor, Sveriges Geologiska Undersökning, Stockholm 50. (1938).
*Bache, Laura, lektor. Eiksbo, vei 509, Eiksmarka, Røa, Oslo. (1931).
Baker, Donald, B.Sc. The Ohio Oil Company, P. O. Box 269, Littleton, Col., U.S.A. (1951).
Bakken, Asbjørn, lektor. Voss Landsgymnas, Voss. (1954).
*Barth, Tom, F. W., professor. Geologisk museum, Oslo NØ. (1921).
Bergersen, Birger, statsråd. Anton Schjøthsgt. 13, Oslo. (1921).
Bergstøl, Sveinung, cand. mag. Geologisk museum, Oslo NØ. (1959).
Bertheau-Hansen, Chr., bergingeniør. Norges geologiske undersøkelse, Oslo NV. (1942).
Birkeland, Tor, lektor. Avaldsnesgt. 36, Stavanger. (1946).
Bjerrum, Lauritz, direktør. Norges geotekniske institutt, Blindern, Oslo. (1952).
*Bjørlykke, Harald, direktør. Norges geologiske undersøkelse, Oslo NV. (1923).
Boughton, Eric. Hay Tor, 921 Harrow Rd., Wembley, Middlesex, England. (1932).
*Broch, Olaf Anton, statsgeolog. Norges geologiske undersøkelse, Oslo NV. (1920).
*Bruun, B., ingeniør. Geologisk museum, Oslo NØ. (1938).
Bryhni, Inge, cand.real. Geologisk institutt, Universitetet i Bergen. (1955).
*Bryn, Knut Ørn, statsgeolog. Borgenvn. 39, Vindern, Oslo. (1956).
Brøndbo, Knut, bergingeniør. Orkla gruber, Løkken Verk. (1955).
*Braastad, Johan, dr. Postboks 808, Oslo. (1913).
Bugge, Arne, dr. Utsiktsveien 21, Stabekk. (1914).
*Bugge, Carl, direktør, Håkon den gode's vei 12, Vindern, Oslo. (S).

*Bugge, Jens A. W., professor. Geologisk Institutt, N.T.H., Trondheim. (1940).

*Bulman, O. M. B., professor. Sedgwick Museum, Cambridge, England. (1951).

Bølviken, Bjørn, ingeniør. Statens Råstofflaboratorium, Postboks 1078, Trondheim. (1958).

*Callisen, Karen, dr. Mineralogisk og Geologisk Museum, Østervoldgade 7, København K. (1917).

Cameron, Richard. Dept. of Geology, Ohio State Univ., Columbus 10, Ohio, U.S.A. (1956).

*Carstens, Harald, geolog. Saxenborg, Trondheim. (1949).

Caster, Kenneth, professor. Dept. of Geology and Geography, University of Cincinnati, Cincinnati, Ohio, U.S.A. (1953).

*Chapman, Charleton, A., professor. University of Illinois, Urbana, Ill., U.S.A, (1949).

*Chen Kwan Yuan, B. Sc. Geological Institute, National Peking University, Peking, Kina. (1951).

Christie, Olav, cand. real. Geologisk museum, Oslo NØ. (1959).

Collini, Bengt, fil.lic., laborator. Kungl. Lantbrukshögskolan, Institutionen för Marklära, Uppsala 7, Sverige. (1949).

*Crowder, Dwight, dr. Mineral Deposits Branch, U.S. Geological Survey, Denver Federal Center 25, Denver, Colorado, U.S.A. (1958).

*Dahl, Eilif, cand. real. Botanisk institutt, Norges Landbrukshøgskole, Vollebekk. (1946).

Dahl, Erling, ingeniør. Skogveien 1, Notodden. (1945).

Dahlström, Elis, dr. Lundberg Explorations Ltd. AB, Lindhagensgatan 132. Stockholm K., Sverige. (1945).

*Danielsen, Anders, konservator. Botanisk museum, Bergen. (1949).

*Devore, George W., dr. Dept. of Geology, University, Chicago 37, Ill., U.S.A. (1948).

*Dietrich, R. V., professor. Dept. of Geological Sciences, Virginia Polytechnic Institute, Blacksburg, Virginia, U.S.A. (1959).

Dietrichson, Brynjulf, bergingeniør. Institutt for geologi, Blindern, Oslo. (1917).

Divljan, Stevan, dr. Federal Geol. Institute ul. Rovinjskaja 12, Beograd, Jugoslavia. (1959).

Donnay, J. D. H., professor. Johns Hopkins University, Baltimore 18, Maryland, U.S.A. (1937).

*Dons, Johannes A., førstekonservator. Geologisk museum, Oslo NØ. (1941).

Dybdahl, Ivar, bergingeniør. A/S Titania, Hauge i Dalane. (1958).

*Eckermann, Harry von, professor. Edeby pr. Ripsa, Sverige. (1937).

*Edelman, Nils, dr. Bolidens Gruv AB., Boliden, Sverige. (1956).

Egeberg, F. P., bergingeniør. Rådhusgata 5b, Oslo (1939).

Einevoll, Ola, landbrukskandidat. Norges Landbrukshøgskole, Vollebekk. (1957).

*Elders, Wilfred Allan, B. Sc. Dept. of Geology, Kings College, Newcastle, England. (1958).

Elliott, Robert B., dr. Department of Geology, The University, Nottingham, England (1954).

Enkovaara, Antti, magister. Oulainen, Finnland. (1947).
Eriksen, Edv. Andreas, bergingeniør. Munkedamsvn. 81c, Oslo (1945).
Erikson, Trygve, chefsgeolog. c/o A. Johnson & Co., Stureplan 3, Stockholm 7, Sverige. (1959).
*Eskola, Pentti, professor, Min.-geol. Inst., Universitetet, Helsingfors. (1919).
Espeland, Margaret, B. Sc. Hus 2, Vei 3672, Nordstrandhøgda, Oslo (1950).
*Falkenberg, Otto, ingeniør. Priorveien 7, Smestad, Oslo. (1914).
Faust, G. T., dr. U.S. Geological Survey, Washington 25 D.C., U.S.A. (1956).
*Feyling-Hanssen, Rolf, statsgeolog. Norges geologiske undersøkelse, Oslo NV. (1950).
*Fuster, José M., professor. Museo National de Sciencias Naturales, Madrid, Spania. (1957).
*Fægri, Knut, professor. Botanisk museum, Bergen. (1935).
Færden, Johannes, statsgeolog. Norges geologiske undersøkelse, Oslo NV. (1948).
*Føyn, Sven, lektor. Sørkedalsveien 93d, Oslo — Smestad. (1932).
Garthus, Torbjørn, lærer. Garthus, Valdres. (1943).
Gavelin, Sven, professor. Mineralogiska Institutet, Stockholms Högskola, Stockholm, Sverige. (1954).
Geis, Hans-Peter, dr. Geofysisk Malmleting, Trondheim. (1958).
Gillberg, Gunnar, fil. dr. Hornsgatan 33 B, Stockholm Sö, Sverige. (1952).
*Gjelsvik, Tore, dr. Evjebakken 18, Sandvika. (1945).
*Gjems, Odd, amanuensis. Norges Landbrukshøgskole, Vollebekk. (1958).
Gjessing, Just, amanuensis. Geografisk institutt, Blindern, Oslo. (1953).
Gjessing, Leiv, læge. Engelsrud, Asker. (1945).
*Gleditsch, Chr., geolog. Kløftaveien 15 B, Hauger, Sandvika. (1941).
Graff-Petersen, Poul, mag.scient. Mineralogisk Museum. Østervoldgade 7, København K. (1957).
*Green, John C. dr. Department of Geology, University of Minnesota, Duluth 11, Minn., U.S.A. (1954).
Grenness, Johannes, ingeniør. Statens råstofflaboratorium, Østmarka, Trondheim. (1945).
*Grip, Erland, fil. lic. Boliden, Sverige. (1938).
*Grønlie, Arne, rektor. Mo komm. høgre almenskole, Mo i Rana. (1943).
Hafsten, Ulf, dosent. Botanisk Museum, Bergen. (1957).
*Hagemann, Fredrik, statsgeolog. Norges geologiske undersøkelse, Oslo NV. (1951).
*Hagner, A. F., professor. Dept. of Geology, University of Illinois, Urbana, Ill., U.S.A. (1947).
Halfdansen, Olaf M., tegner. Grønseth & Co., Skøyen, Oslo. (1958).
Hamar, Georg, vit.assistent. Institutt for Geologi, Blindern, Oslo. (1958).
Hansen, Kaj, dr. Gammeltoftsgade 16, København K. (1947).
*Harry, W. T., B. Sc. Department of Geology, The University, St. Andrews, Fife, Scotland. (1949).
*Hartley, Jack, B. Sc. Department of Geology, University, Leeds 2, England. (1949).

*Hawkes, Leonard, professor, D. Sc., F.R.S. 26 Moore Lane, Rickmansworth, Herts., England. (1915).
*Heber, Gustav, advokat. Oscarsgt. 49, Oslo. (1937).
*Heber, Gustav, bygningsingeniør. Eikskollen 11, Røa, Oslo. (1942).
*Heier, K. S., univ. stip. Geologisk museum, Oslo NØ. (1952).
*Heintz, Anatol, professor. Geologisk museum, Oslo NØ. (1926).
*Heintz, Natascha, vitensk. assistent. Geologisk museum, Oslo NØ. (1951).
Heltzen, Anders M., bergingeniør. Fjellsprengningsutvalget, Løkkeveien 11, Oslo. (1941).
*Helverschou, Kaare, disponent. Villavegen 9, Frøen, Oslo. (1943).
*Henderson, Donald M., professor. Dept. of Geology, University of Illinois, Urbana, Ill., U.S.A. (1949).
*Henningsmoen, Gunnar, førstekonservator. Geologisk museum, Oslo NØ. (1941).
*Hernes, Ivar, amanuensis. Geologisk institutt. Bergen. (1947).
*Hesse, P. A. Jürgen, bergingenjör. Björkhagsvägen, Ormsta, Vallentuna, Sverige. (1936).
*Hessland, Ivar, professor. Stockholms Högskola, Stockholm. (1947).
*Hjelle, Audun, cand. real. Norsk Polarinstitutt, Observatoriegt. 1, Oslo. (1958).
Hjelmquist, Sven, professor. Geologiska Inst., Lund, Sverige. (1949).
Hofseth, Arne, professor. Norges Tekniske Høgskole, Trondheim. (1952).
Holmberg, Sigurd, amanuens. Van Dürens väg 35 A, Lund, Sverige. (1953).
Holmesland, P. B., overingeniør. Herøya pr. Porsgrunn. (1954).
Holmsen, Bjarne, ingeniør. Charlotte Andersens vei 28, Smestad, Oslo. (1941).
*Holmsen, Gunnar, statsgeolog. Skådalen, Oslo. (1908).
*Holmsen, Per, statsgeolog. Norges geologiske undersøkelse, Oslo NV. (1940).
Holtedahl, Hans, førstekonservator. Geologisk institutt, Bergen. (1948).
*Holtedahl, Olaf, professor. Geologisk institutt, Blindern, Oslo. (1908).
*Howe, Henry, V., professor. School of Geology, Louisiana State University, Baton Rouge, Louisiana, U.S.A. (1953).
Huseby, Fredrik, jernbanegeolog. Norges Statsbaner, Storgaten 33, Oslo. (1951).
Hysingjord, Jens, cand. mag. Institutt for geologi, Blindern, Oslo. (1959).
*Høeg, Ove Arbo, professor. Farmasøytisk Institutt, Blindern, Oslo. (1924).
Høegh-Omdal, S. K., byråsjef. Utgarveien 12, Jar pr. Oslo. (1945).
Ihle, Ole-K., cand. real. Norges geologiske undersøkelse, Oslo. NV. (1956).
Ingvaldsen, Karl, dir. Sverresgt. 4, Trondheim. (1956).
*Isachsen, Fridtjov, professor. Geografisk institutt, Blindern, Oslo. (1929).
Iversen, Anker, preparant. Geologisk institutt, N.T.H., Trondheim. (1953).
*Jakhelln, Johanne Hødal, cand. real. United Nations, Box 20, Gr. Central P. O., New York 17, N.Y., U.S.A. (1943).
Jakhelln, Kirsten, fru. Irisvegen 18, Berg, Oslo. (1953).
Johansen, Erling, stipendiat. Kråkerøy, Fredrikstad. (1953).
Jørstad, Finn, geolog. Norges geotekniske institutt, Blindern. (1957).
*Jøsang, Ottar, mag. scient. Geologisk Institutt, N.T.H., Trondheim. (1955.)
*Kautsky, Gunnar, statsgeolog. Sveriges Geologiska Undersökning, Stockholm 50, Sverige. (1952).

*Kennedy, W. Q., D. Sc., F. R. S., professor. University of Leeds, Leeds 2, England. (1947).
*Kiil, Erling, bergingeniør. Brantenborgveien 2, Slemdal, Oslo. (1922).
*Klompé, Theo, professor. Dept. of Geology, Faculty of Science, Chulalongkorn University, Phya Thai Rd., Bangkok, Thailand. (1947).
Klüver, Emil, ingeniør. Statens toll-laboratorium, Oslo. (1926).
*Kolderup, Niels-Henr., professor, Geologisk institutt, Bergen. (1919).
*Kollerud, Martha, cand. mag. Trondheimsveien 10, Oslo. (1919).
Kraft Johanssen, J., direktør. A/S Sydvaranger, Kongens gt. 5, Oslo. (1955).
*Krauskopf, K. B., professor. Geol. Dept. Stanford University, Stanford, Cal. U.S.A. (1953).
*Kristoffersen, K., dr. A/S Norsk Jernverk, Mo i Rana. (1941).
*Kullerud, Gunnar, dr. Geophysical Laboratory, 2801 Upton Street, Washington 8, D.C., U.S.A. (1948).
*Kulling, Oskar, statsgeolog. Sveriges Geologiska Undersökning. Stockholm 50, Sverige. (1953).
Kvale, Anders, professor. Geologisk institutt, Bergen. (1936).
Kvalheim, Aslak, laboratoriesjef. Statens råstofflaboratorium, Østmarka, Trondheim. (1944).
Landergren, Sture, dr. Sveriges Geologiska Undersökning, Stockholm 50. (1943).
Landmark, Einar, direktør. Tidemandsgt. 14, Oslo. (1941).
Landmark, Kåre, direktør. Tromsø museum, Tromsø. (1943).
*Larssen, Kari Egede, statsgeolog. Norges geologiske undersøkelse. Oslo NV. (1955).
*Larsson, Tollef, cand. real. Engene Fabrikker, Sætre i Hurum. (1945).
Laursen, Dan., lektor. Julius vej 6, Gentofte, København. (1947).
*Lien, Håkon, geolog. Suite 2001, 80, Richmond St. W., Toronto, Ontario, Canada. (1955).
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2	korresponderende medlemmer.
130	livsvarige medlemmer.
127	årsbetalende medlemmer.
259	medlemmer.